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REFERENCE

Lane, B. H., 1935, Suggestions to authors of papers submitted for publication by the United States Geological Survey: U. S. Geol. Survey, Washington, D. C. (For sale by Superintendent of Documents, Washington, D. C. 40c.)

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MARCH 1957

ARROYOS AND THE SEMIARID CYCLE OF EROSION

S. A. SCHUMM and R. F. HADLEY

ABSTRACT. Longitudinal valley profiles, surveyed in small drainage basins in eastern Wyoming and northern New Mexico, indicate that discontinuous gullies are often developed on areas of local steepening of valley fills.

The large amounts of runoff lost through channel absorption in ephemeral streams cause increased sediment loads downstream. This promotes aggradation within these valleys and eventual dismemberment of the drainage system by the sealing off of tributary channels. Continued aggradation in these valleys steepens the gradients, causing the formation of discontinuous gullies and reintegration of the system by arroyo cutting in the fills.

INTRODUCTION

Arroyos or gullies trenching valley fills are common throughout the semiarid regions of the West. The cause of arroyo cutting has interested many investigators, mainly because of the economic losses resulting from their formation.

Current hypotheses of the origin of arroyos have been succinctly summarized by Antevs (1952). Briefly, two schools of thought are dominant: (1) Depletion of vegetative cover by overgrazing is the cause of present arroyo cutting (Bailey, 1935; Thornthwaite, Sharpe, and Dosch, 1942; Antevs, 1952); (2) Overgrazing may be an initiating factor, but the chief cause is a climatic change (a) to the drier (Bryan, 1925), (b) to the more humid (Dutton, 1882; Barrell, 1908; Huntington, 1914; Gregory, 1915, 1917; Bryan, 1922, p. 85), (c) either a change to the drier or to the more humid (Richardson, 1945), or (d) a change in rainfall intensities (Leopold, 1951).

It is generally agreed that the introduction of large numbers of cattle began about 1870 and that arroyo cutting became serious beginning about 1880 (Antevs, 1952). Arroyo cutting following overgrazing seems a logical conclusion.

It is difficult to evaluate the possibility of a change in climate because of the few meteorological records available. Trends have been found in some data, but even a proved trend in 50 to 100 years of record may be little more than a temporary fluctuation from wet to dry conditions during that period. This is amply demonstrated by the results of tree-ring analysis. Analysis of many hundreds of years of tree-ring variations, although in themselves revealing little about long-term climatic change, indicate that a rainfall record of 100 years duration cannot be cited as evidence for a climatic change (Schulman, 1954).

Schulman's studies of tree-ring chronologies for the Upper Colorado River (1945), Mesa Verde, Gila, and southern Arizona regions (1942), as well as the correlation of tree growth with known rainfall and runoff records, afford information on fluctuations in tree growth and rainfall for a period

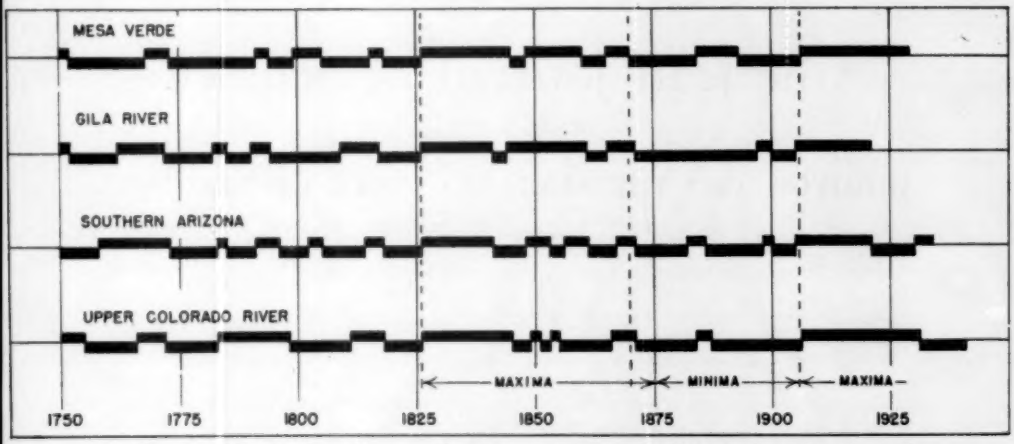

Fig. 1. Periods of maximum and minimum tree growth. The portion of each graph above the mean line indicates years of maximum tree growth, whereas that portion below the mean line indicates years of minimum growth.

from A. D. 1298 to 1940. Schulman has prepared tables indicating periods of maximum and minimum tree growth. For the purposes of this paper, Schulman's data for the period 1750 to the present have been plotted in figure 1. The periods of maximum and minimum tree growth as plotted do not indicate the severity of each period, but as Schulman suggests the relationship of tree growth to precipitation indicates that in a general way these periods are closely related to maxima and minima of winter precipitation. It may be that the years of maximum tree growth are also years of good grass cover.

The plots of figure 1 may be divided into four segments: the period from 1750 to 1826 is in general a period in which minimum growth exceeds growth above normal; the period from 1826 to 1870 is clearly a period of maximum growth, except in the upper Colorado River region where the period is shorter, ending about 1855. The period 1870 to 1906 is a period of marked minimum growth, whereas after 1906 a period of maximum growth is maintained until between 1921 and 1931.

The importance of these periods is obvious, for the lush vegetation in many valleys, described by the early travelers in the Southwest, may be correlated with the period of maximum tree growth of 1826 to 1870. Coincident with the introduction of large numbers of cattle in the Southwest in 1870 a period of minimum growth began, interrupted by short periods of higher rainfall from 1881 to 1884 (Peterson, 1950) and higher intensity rains (Leopold, 1951).

It seems, then, that each of the hypotheses listed above has some validity, for if vegetation was depleted it was depleted both by the cattle and by dry conditions associated with a period of minimum tree growth. To add to the complexity of the problem, arroyo cutting began in some areas before they were grazed, and in other heavily grazed areas no arroyos have developed (Peterson, 1950, p. 421).

It is the purpose of this paper to present data, collected in small drainage areas, suggesting the origin and location of discontinuous gullies within these valleys may be dependent on the gradient of the valley fill, and that any hypothesis of arroyo cutting must take into consideration the character of the drainage basin itself.

The writers' observations in the Cheyenne River basin above Angostura Reservoir, Wyoming, South Dakota, and Nebraska and in Sandoval and San Juan Counties, New Mexico, are the basis of the following discussion. Because the studies were made in small valleys tributary to Lightning Creek (Cheyenne River basin), Arroyo Torreon (Rio Puerco drainage system) and the Mancos River a direct comparison should not be made between the present studies and longer arroyos typified by the Rio Puerco.

THE ORIGIN OF DISCONTINUOUS GULLIES AND ARROYOS

Observation in Wyoming and New Mexico.—In order to obtain quantitative information on the characteristics of a drainage system that might influence arroyo cutting, longitudinal profiles of several trenched valleys were surveyed.

Because alluviation within a valley steepens its gradient, it was assumed that the alluvial fill might be built up to a critical angle and then trenched. To determine if this were true, it was necessary to survey the profiles of both the present channel and the surface of valley fill adjacent to the trench.

Trenched reaches of several small valleys were surveyed in Wyoming and New Mexico. In each case the valley contained what appeared to be a modern alluvial fill, for although vegetation was present no distinct soil profile had developed on the fill surface, and recent artifacts were found in two of the fills. The alluvium generally was derived from one gently dipping lithologic unit, and, therefore, was of nearly uniform texture and composition where sampled within the valleys.

In none of the surveyed gullies did bedrock influence the profiles to any noticeable extent. None of the headcuts or breaks in gradient were caused by the appearance of bedrock in the channel. In some cases, bedrock was exposed in the gully floors, but its effect was probably limited to a retardation of incision or a tendency toward lateral planation. No evidence of either was observed.

The profile shown in figure 2 is that of a portion of Manning Draw located in sec. 36, T. 35 N., R. 67 W., Niobrara County, Wyo. The drainage area above the surveyed reach is 1.7 square miles. The valley fill, derived from the Paleocene Fort Union formation is composed of 45 percent clay-size particles, 34 percent silt, and 21 percent sand.

A marked convexity exists on the fill surface above the lower gully in figure 2. The convexity is not as obvious near the mouth of the upper gully, but the gradient of the fill is definitely steeper at the lower end. At the mouth of the lower trench the gradient of the valley fill is 0.027 while the floor of the trench has a gradient of only 0.017. Upstream from the mouth of the trench the gradient of the fill increases to 0.036 and then gradually decreases to 0.009 at the upper extremity. The presence of both of these gullies appears

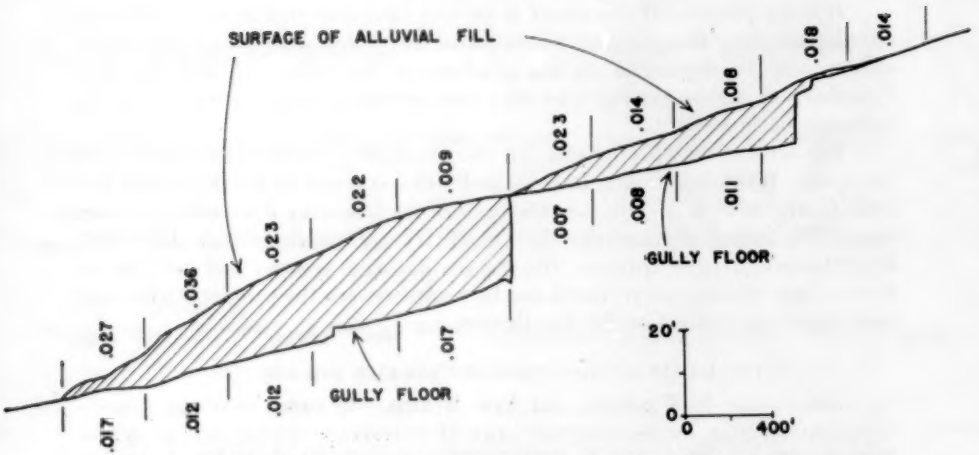

Fig. 2. Profile of discontinuous gullies in Manning Draw, Niobrara County, Wyo. Figures are gradient for each section of the profile in ft./ft.

to be due to a localized steepening of the alluvial fill. Trenching was probably initiated on the steeper sections of the fill, 0.036 on the lower and 0.023 on the upper gully.

If trenching began on the section with a gradient of 0.036 at the lower end of the fill, then the gully has apparently extended itself some distance downstream from the point of initial trenching, contemporaneous with upstream headcut migration. Further headward development of the lower gully will undoubtedly cause integration of the two trenches.

Harney Creek Draw, with a drainage area of about 1.25 square miles, located in sec. 22, T. 34 N., R. 67 W., is another example of a trenched valley, developed in alluvium derived from the Fort Union formation, showing relationships similar to those on Manning Draw (fig. 3A). Only one gully is present in the valley, but again the steepest part of the alluvial fill surface, in the lower reach of the valley, is located just above the mouth of the gully, suggesting that it was the point of initial cutting. Upstream from this point the gradient on the fill is 0.008 and downstream it is 0.004. Farther upstream both the gradient of the gully and the fill surface increase toward the drainage divide.

Joe Warren Draw, located in sec. 34, T. 35 N., R. 67 W., on the Fort Union formation, contains three discontinuous gullies within a drainage area of 0.6 square miles (fig. 3B). The mouth of each is located at a change in gradient of the fill surface. The gradient of the fill at the downstream end of the lower trench is 0.015 which decreases to 0.009 upstream until the second trench appears where the gradient is 0.013. At the upper end of the valley a small trench has begun to cut in a reach where the gradient steepens to 0.016. The middle of the three trenches shows two sections of steeper gradient, marked by a convexity of the fill surface at each point. The presence of a knickpoint in the profile of the trench suggests that two discontinuous gullies

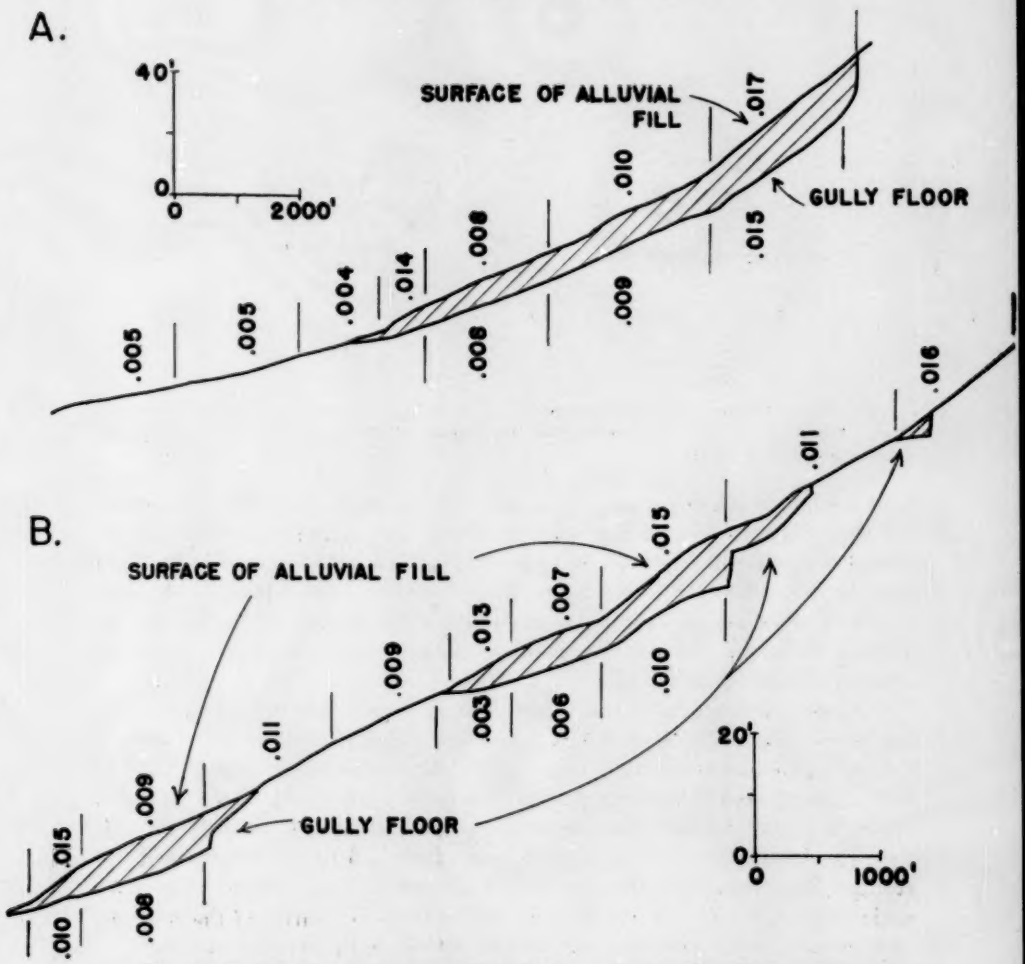

Fig. 3. Profiles of discontinuous gullies, Niobrara County, Wyo. Figures are gradient for each section of the profile in ft./ft.

A. Harney Creek Draw

B. Joe Warren Draw

developed on the steeper segments of the fill, which were united by the headward growth of the lower trench.

Traphagen Draw is located in sec. 5, T. 37 N., R. 64 W., Niobrara County, in alluvium derived from the Upper Cretaceous Lance formation. Profiles A and B of figure 4 reveal the abrupt changes in gradient and a reversal of gradient at the mouth of two gullies due to deposition and the formation of an alluvial fan. The integration of the headward eroding downvalley arroyos with the upper gullies is hindered by the high rates of deposition between the two. It is possible that the mouth of the upper gullies will migrate

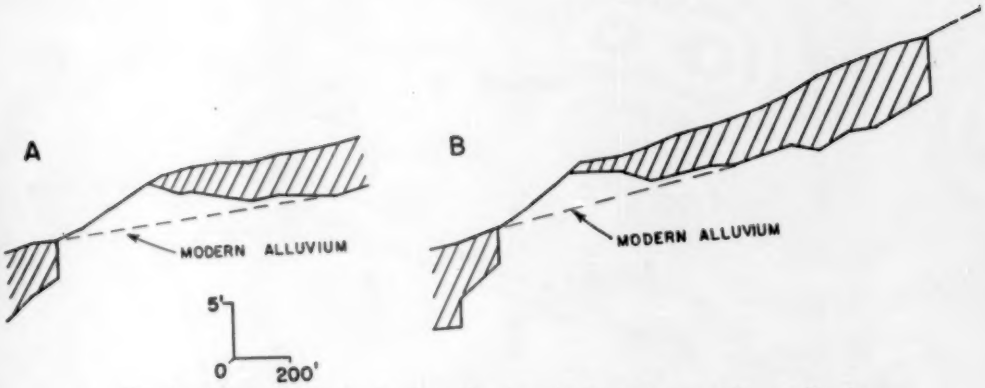

Fig. 4. Alluviation in the lower reach of small discontinuous gullies in Traphagen Draw, Niobrara County, Wyo., hindering integration of the upper and lower gully and causing reversal of gradient.

upstream as the lower reaches of the gully are buried by deposition, maintaining their distance between the lower headcuts. This recent deposition has obscured the point of origin of the upper gullies. It has been noted in the field that the fill surface, adjacent to the longer arroyos, shows many such areas of recent overbank deposition, which obscure the history of the arroyo by burying points of probable origin of discontinuous gullies or creating other areas of steepening on the fill.

A small, well-vegetated valley underlain by Pierre shale of Late Cretaceous age in sec. 15, T. 36 N., R. 63 W., Niobrara County, contains three discontinuous gullies and intervening fans (fig. 5). The entire sequence is only 1,200 feet in length and the drainage area above the upper gully is less than 0.5 square mile, yet similar slope conditions prevail as were found in larger valleys. The headcuts occur on slopes ranging from 2.5 to 3.5 percent and the average slope between gullies is only 1.5 percent. The measurements were made using a hand level, but they do demonstrate the clarity of the relationships between steep gradients and trenching even in the smallest of valleys.

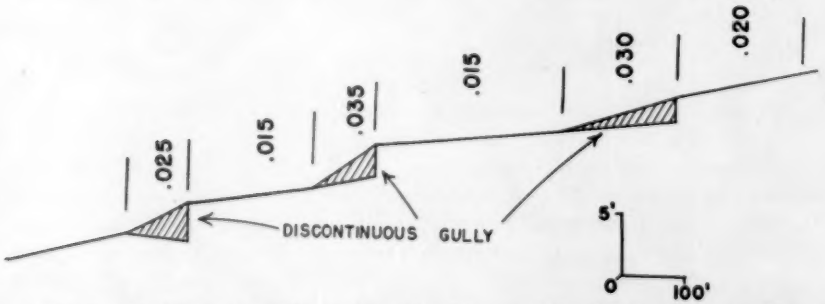

Fig. 5. Discontinuous gullies in a small well-vegetated valley in Niobrara County, Wyo. Figures are gradient of profiles in ft./ft.

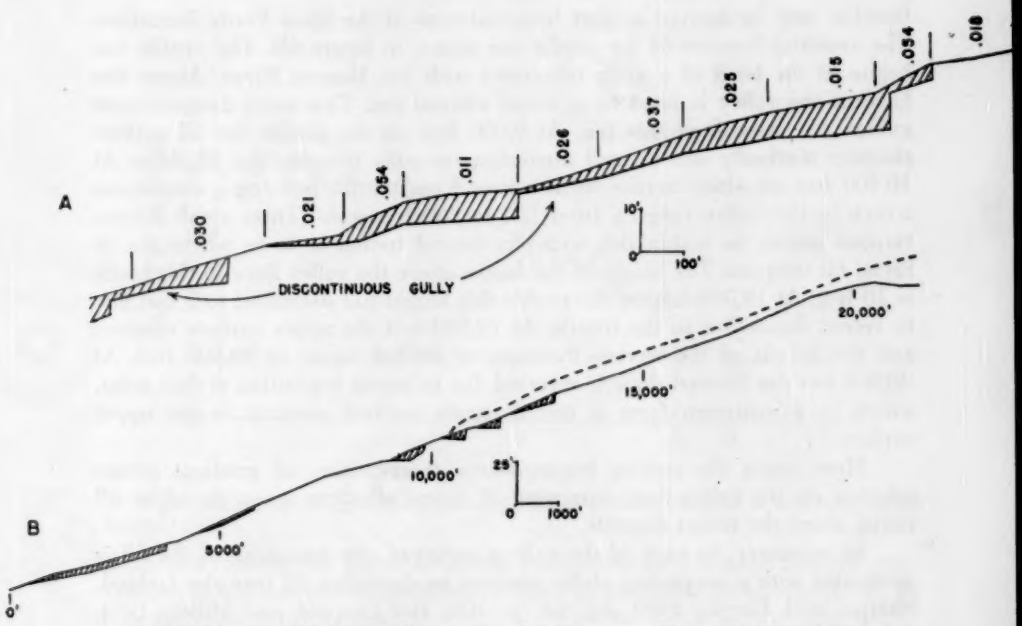

Fig. 6. A. Discontinuous gullies in Dam 17 Wash near Cuba, N. Mex. Figures are gradient of profile in ft./ft.

B. Longitudinal profile of tributary to the Mancos River, San Juan County, N. Mex. Dotted line is profile of older terrace surface. Hatched areas are discontinuous gullies.

Figure 6A shows a reach of channel surveyed in Cornfield Wash near Cuba, New Mex. in a drainage area of about 0.5 square mile. Two discontinuous gullies trench the alluvium, which is derived from a predominantly shale member of the Mesa Verde formation of Late Cretaceous age. Grain-size analysis of a composite sample of the fill shows: 30 percent clay-size material, 54 percent silt, 12 percent sand, and a 4 percent fine to medium gravel. The relationships in the gully farthest downstream on this profile are obscured by overbank deposition caused by a small dam near its mouth. However, each of the gullies appear to be formed by the union of two smaller discontinuous trenches. Where irregularities on the valley fills are closely spaced the integration of small gullies may be common, forming larger more efficient channels. In addition, the profile shows that the relationships observed in Wyoming are similar to those in New Mexico with the steepest parts of the profiles occupied by discontinuous gullies.

To study the relationships in larger valleys an area in New Mexico just south of the Colorado State line was visited, and a reach in a tributary to the Mancos River with a drainage area of about 19 square miles was surveyed. The valley-fill was derived predominantly from the Mancos shale of Late Cretaceous age. A mean of ten samples of the fill indicate that it is composed of 27 percent clay-size particles, 57 percent silt, and 16 percent sand. The sand

fraction may be derived in part from outcrops of the Mesa Verde formation. The essential features of the profile are shown in figure 6B. The profile was begun at the head of a gully integrated with the Mancos River. Above this headcut the valley is filled by a broad alluvial fan. Two small discontinuous gullies are present on this fan. At 9,000 feet on the profile the fill surface steepens markedly and a small discontinuous gully trenches the fill there. At 10,400 feet an older terrace surface rises from the fill forming a continuous trench in the valley ranging from 100 to 200 feet wide. Three small discontinuous gullies lie within this wide flat-floored trench in areas where the alluvial fill steepens. The height of the banks above the valley floor in this reach is 10 feet. At 19,000 feet on the profile this height has decreased to 5 feet due to recent deposition in the trench. At 19,500 feet the upper surface steepens and the height of the terrace increases to 10 feet again at 20,500 feet. At 20,500 feet the channel slope is reversed due to recent deposition at that point, which is downstream from a recent trench actively eroding in the upper surface.

Here again the cutting begins where a steepening of gradient occurs whether on the valley floor composed of recent alluvium or on the older fill rising above the recent deposits.

In summary, in each of the valleys surveyed, the trenching of the fill is associated with a steepening of the gradient on the valley fill (see also Ireland, Sharpe, and Eargle, 1939, fig. 63, p. 100, and Leopold and Miller, 1954, fig. 25, p. 81). In some cases the cutting appears to have originated on these steeper reaches and the channel has eroded up-valley with little down-valley development (figs. 2 and 5). In other areas the gullies have extended themselves down-valley also (figs. 3 and 6A). In some valleys deposition at the lower end of the trenches appears to prohibit integration with other gullies (fig. 4) whereas in other valleys integration has apparently occurred or will occur (figs. 2, 3B, and 6A).

In the larger valleys studied in eastern Wyoming and New Mexico, ranging in size from 0.6 to 19 square miles cutting is begun on slopes of from 1.5 to 2.5 percent, whereas in smaller valleys ranging in size from 0.05 to 0.5 square mile the steeper portion of the fill is 2.5 to 5.4 percent.

The increase of the angle at which trenching occurs with decreasing size of the drainage basin strongly suggests an inverse correlation of angle at which the fill will be trenched or instability of the fill with discharge; i. e., for a given storm, discharge will be less from the small drainage systems and cutting will not occur on a slope as flat as it will in a neighboring larger basin with greater discharge. This may also explain the ease of recognition of steepened portions of fills in the smaller valleys.

The decreasing stability of the fill during alluviation, caused by increased gradient, may be expressed graphically as done by Terzaghi (1950) for earth slopes prior to failure (fig. 7). Superimposed on a descending curve of decreasing stability are many minor fluctuations caused by high intensity rains, floods, or depletion of vegetation. The effect of these fluctuations on the fill is minor until the factor of safety or stability of the fill has been so reduced by oversteepening that during one fluctuation the arroyo cutting begins. It is

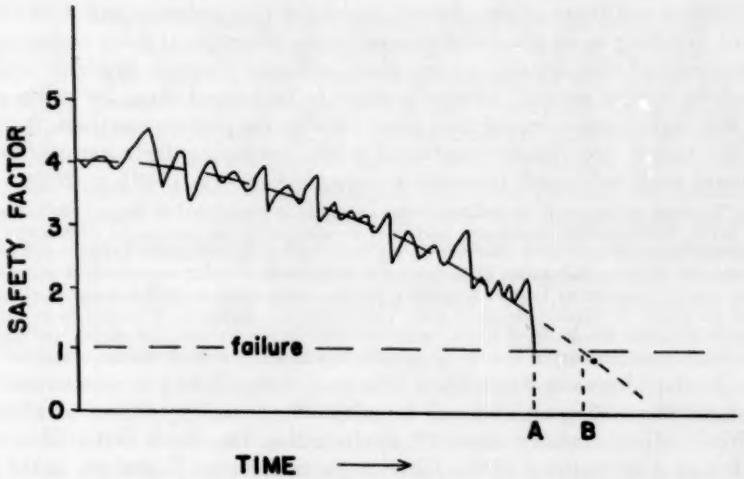

Fig. 7. Possible changes in the safety factor of a valley fill prior to trenching (after Terzaghi).

important to note that the cause of the fluctuation, be it a large flood or overgrazing, is not necessarily the primary cause of the valley trenching; it may be only the most obvious cause.

Restated in another fashion, on a slope of a certain inclination trenching will not occur at a given discharge, Q_1, until additional deposition steepens the fill. If floods of discharge Q_1 were of the greatest magnitude possible, then trenching would be delayed until time B (fig. 7). However, if a flood of doubled discharge occurs, Q_2, due perhaps to a storm of high intensity then cutting might occur at time A, for the increased mass of water and velocity due to the greater discharge (Q_2) will initiate cutting where it could not occur under the conditions of discharge Q_1. This concept appears to agree with the field observations of a steeper angle at which cutting begins in the smaller valleys.

Observation made by other writers.—Some examples of similar cyclic alternation of erosion and sedimentation in stream valleys have been observed which illustrate that these alternations may be a natural part of the cycle of erosion. These are enumerated below.

Thornthwaite, Sharpe, and Dosch (1942, p. 89) clearly suggest this possibility in the Polacca Wash:

> " . . . . it is believed that the explanation of successive deposition and removal of fill lies in sedimentary processes and irregular occurrence of heavy storms rather than change of climate."

McGee (1891, p. 261), referring to work in Iowa, stated:

> "One who lives in a region of delicate adjustment between water work and configuration, such as the driftless area, cannot fail to observe the alternate formation and destruction of miniature terraces in every minor valley or ravine within a period ranging from a year to a generation."

According to McGee, the valleys are plugged by tributaries building fans or deltas into the main valley or the dropping of stream load for any reason.

Deposition continues in the channel above the plugged zone and works headward, resulting in an alluviated channel above the original point of deposition. Trenching of this deposit occurs when sufficient water is available and the gradient is steep enough, causing terraces to be formed along the valley sides. In this region, more humid than those cited in the previous sections, the walls of the trench are rapidly weathered back, destroying their arroyo-like apperance until only small remnants remain. McGee says (1891, p. 262):

> "In some quite small periodic streams the cycle is completed in five or ten years, and the birth, development, decadence, and final dissolution of the successive straths and their resulting terraces have been observed in the same valley. In the larger periodic and smaller permanent streams the stages of half a cycle have been noted, progressing at such a rate as to require quarter or half of a century for its completion; in mill-streams single stages (but all stages in various streams) have been observed during such period as to indicate a cycle of some hundreds of years; while in still larger streams the stages are obscure, and the alternation during a score of years is too inconspicuous to attract attention."

In the Cheyenne River basin (Sioux County, Nebr.) a cottonwood tree approximately 60 years old was found partly buried by 8 feet of alluvium within a valley draining about 20 square miles. The depth of burial was revealed by a retrenching of the fill adjacent to the tree. Therefore, in the past 60 years the channel was filled by 8 feet of alluvium and then cut again to its present depth slightly below the roots of the tree. Renewed deposition is occurring at present some distance downstream from the tree. This example suggests that in small valleys the process occurs as visualized by McGee and at rates more rapid than generally believed possible.

SOME ASPECTS OF THE CYCLE OF SEMIARID EROSION

The presence of oversteepened reaches within an alluviated valley, and the appearance of discontinuous gullies on these over-steepenings suggest that some natural cycle of alluviation and erosion is operative in semiarid valleys. The problem still remains as to the reason for alluviation or valley filling. Apparently, it is due to a deficiency of water in relation to sediment.

Hydrologic considerations are, therefore, of major importance in any consideration of stream action in areas of low rainfall. It has been shown (Leopold and Miller, 1956) that sediment load in the arroyos of New Mexico increases faster than discharge. Hadley reports three occurrences of floods in Wyoming in which the discharge was greatly reduced in a downstream direction due to channel absorption in the stream beds (see table 1). In addition, recent hydrologic studies in the Cheyenne River basin have shown that a large percentage of runoff from headwater areas is lost in the channels before reaching a master stream, (Culler, 1956). Sediment being transported in this type of channel must be deposited as the flow is dissipated.

TABLE 1

Downstream Decrease in Discharge Due to Channel Absorption

Date of observation	Creek	Q upstream (est.)	Q downstream (est.)	Distance between points (miles)
July 18, 1953	Twentymile	40 cfs	0	6
Aug. 3, 1953	Twentymile	450 cfs	42 cfs	7
Aug. 16, 1953	Lightning	100 cfs	40 cfs	5

Observations in Wyoming and elsewhere reveal that one other important characteristic of a semiarid drainage system is the lack of accordance of tributary and main channel, that is, the tributaries may or may not be graded to the main channel. As an example of this, the Twentymile Creek drainage basin may be cited where one-third of the tributaries are graded to the surface of a terrace above the main channel (Hadley, 1953). This lack of integration in certain channels is due to the filling of the lower parts of the tributary valleys with sediment, causing spreading of water and sediment over the terraces and flood plain adjacent to the main channels. In any larger drainage system the tributaries may be found in all stages of integration with the main channel, suggesting that each tributary has its own history of alluviation and dissection which may not be contemporaneous with that of its neighbors.

With the above two main characteristics of a semiarid drainage network in mind, it may be possible to hypothesize a cycle of semiarid erosion based on field observations within the Twentymile Creek drainage basin, Wyoming. Twentymile Creek drains an area of about 200 square miles underlain by the Fort Union formation in which mean annual rainfall is about 14 inches. The main channel of Twentymile Creek is tributary to Lightning Creek, and throughout its length has a well-defined channel, i. e., aggradation is minor within the channel.

Field examination of a number of tributaries to Twentymile Creek show in general two cases: (1) actively eroding headwater channels and filled or alluviated lower valley reaches and (2) moderate erosion in headwater channels and gullies trenching the lower valley fill. These two seemingly distinct examples appear to be the two components of the semiarid cycle of valley development. This cycle may be outlined as follows: An alluviated tributary valley is united with the major drainage channel by the development of a trench in the recent alluvium clogging the tributary channel. The gully is extended headward by upstream headcut migration. Figure 8A shows the channel at the beginning of this rejuvenation. As the headcut migrates up channel the lower section of the tributary drainage (sec. 1, fig. 8A) becomes very efficient for sediment transport, for the runoff is concentrated in a clearly defined channel. The headcut continues to work up channel, passing tributaries 2 and 4 and rejuvenating them in turn (fig. 8B). Runoff increases and time of concentration is much shorter, but the volume of sediment moved is greatly

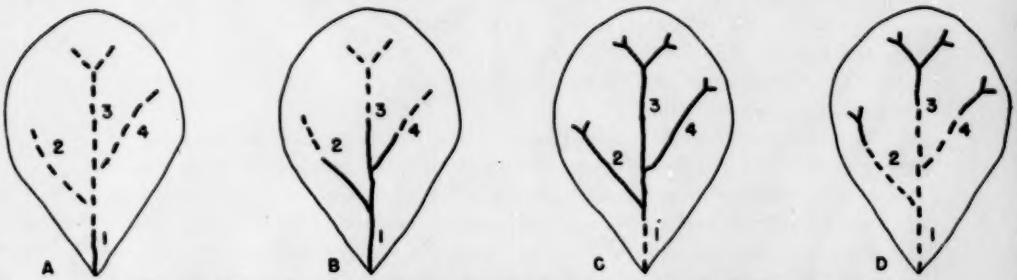

Fig. 8. The cycle of trenching and alluviation in a semiarid valley. Dotted line indicates alluviation within a channel. Solid line indicates trenching or a through channel.

increased by the rejuvenation of tributaries 2 and 4. With the rejuvenation of tributaries the upland slopes may be steepened and they in turn may supply more sediment to the channels. Thus, as the headward cutting continues, up-channel sediment delivered to section 1 of this basin is greatly increased as is the velocity of flow but with perhaps only minor increase in runoff. Sections 2, 3, and 4 now are supplying much sediment, but section 1 has a gradient formed when the supply of sediment was lower and aggradation begins in section 1, perhaps due also to widening of the channel (Leopold and Miller, 1956), or deposition may begin at the junction of tributary with master stream (fig. 8C). Thus, aggradation begins as discontinuous deposits in reaches of the valley with either the least slope or greatest width or both. Alluviation once begun promotes more alluviation until the channel is filled and is no longer smoothly graded to the main drainage channel. The area of maximum channel deposition then migrates upstream. As it passes the mouths of tributaries they begin filling their valleys, for the local base level is raised. The area supplying sediment to the main channel decreases in size (fig. 8D) and the rate of alluviation undoubtedly decreases. The gradient of the filling valley becomes steeper until discontinuous gullies begin to form in stretches of channel where the gradient is steepest. These discontinuous gullies shift the alluvium down channel by building fans in the valley which are in turn trenched when the gradient steepens. Finally the shifting of sediment down-channel forms a fan near the tributary mouth. This fan is in turn trenched, integrating the tributary drainage with that of the master stream, and so the cycle begins again.

This hypothetical cycle explains many of the relationships observed in small valleys in Wyoming, and is a partial development of McGee's statements (1891). Probably this proposed cycle or some modification of it is typical of semiarid channels and even intermittent channels in more humid regions (Happ, Rittenhouse, and Dobson, 1940).

CONCLUSIONS

The profiles of valley fills and discontinuous gullies surveyed in Wyoming and New Mexico strongly suggest that discontinuous gullies form on reaches of steeper gradient within a valley. The angle at which the fill will be trenched is probably dependent on discharge. In two similar valleys, trenching will occur soonest in the valley with the highest frequency of large floods.

Semiarid and probably arid regions are distinguished from the more humid by lack of accordance of many stream junctions and high rates of water loss due to channel absorption, which cause an increase in sediment concentration downstream.

The cycle of erosion within a large drainage system may be made up of major cycles of arroyo cutting on the main channel, but within these major cycles are a number of epicycles in which alluviation and erosion alternate in the smaller valleys, making them temporarily independent of the main drainage channels. The inception of cutting within the smaller valley appears to be dependent on the angle to which the fill has been steepened and the incidence of high intensity precipitation. Since the smaller valleys are found in all stages

of the cycle of aggradation and trenching, these epicycles are probably dependent mainly on local factors.

The possibility that cutting is started on steeper reaches of even the largest valleys might be difficult to prove, although it may be that principles of gully formation discovered in small valleys are applicable to the larger ones.

The minor fluctuations within the cycle then do not markedly affect the major landforms but will cause the development of valley-plug deposits or the formation of terraces (arroyo cutting) in smaller valleys. The epicycles become more prominent in arid and semiarid regions as the readjustment, owing to the deficiency of water, will not occur as rapidly. These epicycles are then minor parts of the main erosion cycle, but they exert a great influence on the economy of semiarid areas.

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HEAVY MINERALS IN ARENACEOUS BEDS IN PARTS OF THE OCOEE SERIES, GREAT SMOKY MOUNTAINS, TENNESSEE*

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ABSTRACT. Arenaceous beds of the Ocoee series (Precambrian?), now largely low-grade metamorphic rocks, in the Great Smoky Mountains of eastern Tennessee contain a restricted suite of heavy minerals dominated by varying proportions of ilmenite and zircon. This study of the heavy minerals was facilitated by using the saprolitic rock. Comparison of mineral concentrates from saprolites with those of fresh rock shows that the major changes are the removal of apatite, alteration of ilmenite, and oxidation and hydration of magnetite. Anatase is authigenically developed from ilmenite during the saprolitization process.

Three principal varieties of zircon were recognized in the heavy residues: (1) purple or hyacinth, well rounded, and fresh; (2) metamict, angular to subangular; and (3) clear colorless, subangular to rounded. The average composition of the zircon population in the 26 samples examined is: (1) 48 percent, (2) 8 percent, and (3) 44 percent. Samples of these three varieties of zircon were purified for age determination by the total lead-alpha (Larsen) method. Ages determined ranged from 620 million years (fresh purple elongated grains) to 1140 million years (fresh purple rounded grains), with metamict and colorless zircon between 740 and 890 million years. Grouping of the age determinations suggested that these arenaceous beds received zircon from several sources.

Granite boulders from a conglomerate in the same sequence yielded zircon unlike that in the arenaceous beds, as did granite and gneiss of the crystalline basement. Arenaceous sequences—a northern and a southern—distinguishable from each other in the field, could not be differentiated by their heavy minerals.

INTRODUCTION

This paper gives the results of a study of the heavy-mineral assemblages of some arenaceous beds (hereinafter referred to as arenites) of the Ocoee series, probably Late Precambrian age (King, 1949), in the Great Smoky Mountains of eastern Tennessee. This investigation forms part of a study of the geology of the Great Smoky Mountains National Park by the U. S. Geological Survey under the direction of Philip B. King. Several members of the field party assisted in collecting the samples; these were examined mineralogically in the sedimentary petrology laboratory at Beltsville, Md. Purification of zircon concentrates and age determinations were made by Howard W. Jaffe. We are indebted to Professor F. J. Pettijohn for a critical examination of these rocks and for advice in nomenclature.

The objectives of the investigation were twofold: (1) to relate the heavy minerals to the units defined in field mapping, and (2) to obtain information about the source or sources of the arenites. Much of the rock examined was soft and weathered, and is known throughout the southeastern United States as saprolite (Becker, 1895). Saprolite retains most of the heavy minerals that are present in the fresh rock.

GENERAL GEOLOGY

The arenites from which the heavy minerals were obtained (via their saprolites) are hard, gray, compact medium- to coarse-grained rocks. Examination of thin sections has identified them as varieties of subarkoses (Pettijohn, 1954, p. 364) which have undergone low-grade metamorphism. This has re-

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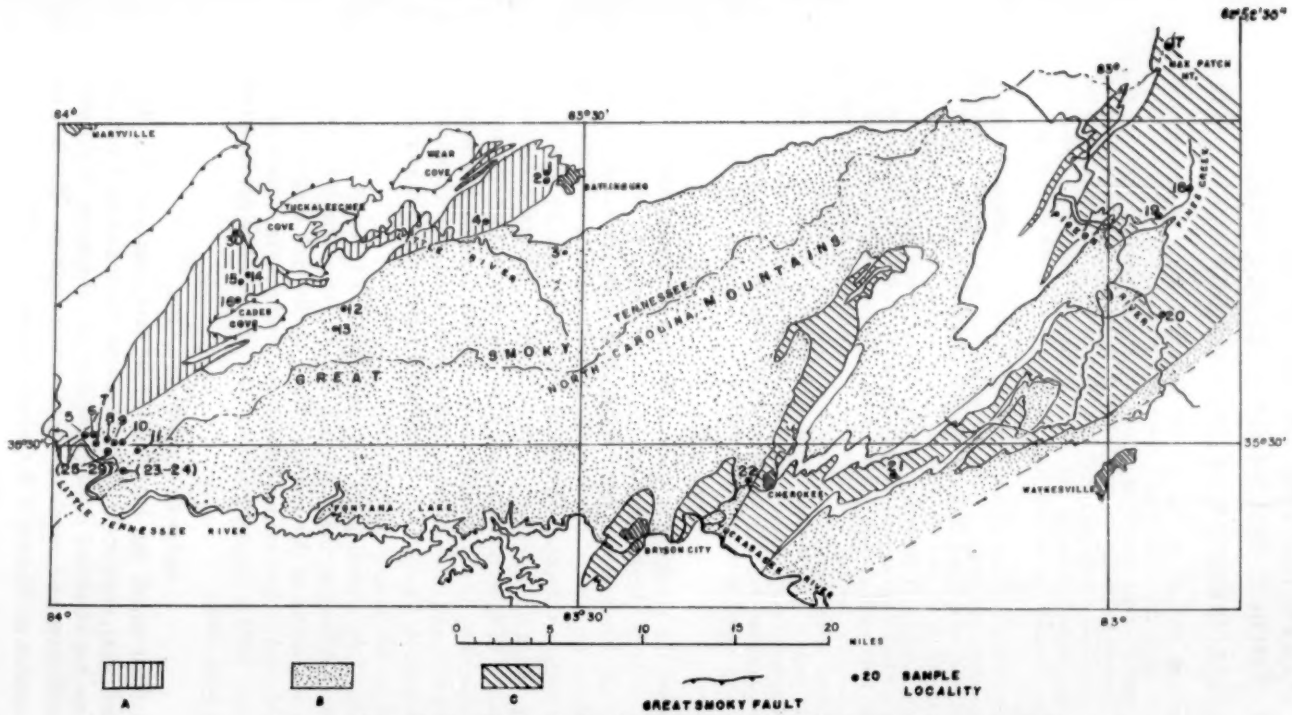


Fig. 1. Map of the Great Smoky Mountains National Park area, Tennessee and North Carolina, showing distribution of pertinent rock units and points from which samples were collected (adapted from Hadley, King, Neuman, and Goldsmith, 1955, pl. 1).

A, Northern arenaceous sequence of Ocoee series
B, Great Smoky conglomerate of Ocoee series, composed largely of southern arenaceous series. } Precambrian(?)

C, Crystalline basement — Precambrian

Blank areas on the map include Snowbird formation and related rocks of Keith, and rock units not discussed in this paper. The Great Smoky fault separates the Ocoee series and related rocks from the Paleozoic rocks of the Tennessee Valley.

sulted in the alignment of quartz and sericite or biotite grains. This effect is not recognizable in hand specimen or in rock outcrop.

The arenite beds are 6 inches to 6 feet thick and commonly show graded bedding. Many beds terminate upwards in argillite, siltstone, or slate partings. Fragments of argillite are common in many of the coarser beds, and argillite occurs throughout the sequence, ranging from thin partings to beds more than 100 feet thick.

In the field a northern and a southern sequence of arenite were mapped (fig. 1). Although the northern and southern sequences are much alike they may be distinguished by minor petrographic differences such as:

(1) Blue quartz is a common constituent of the southern but not of the northern sequence.

(2) Fragments of granitic rocks are common in the northern but not in the southern sequence.

(3) Sorting of the coarser constituents is much poorer in most of the northern sequence. Grains 5 to 10 mm in diameter in coarse-grained sandstone are common in the northern but are rare in similar beds in the southern sequence.

In addition the southern sequence in some places contains spherical bodies, 6 to 12 inches in diameter and a little darker in color than the surrounding rock, which have been called concretions (Stose and Stose, 1949, p. 281, pl. 3, fig. 1; King, Hadley, and Neumann, 1952, p. 49). Mineralogically there was no difference in the concretion and its surrounding rock at the one locality examined in detail.

In thin section the arenites show considerable variation in texture, due in part to grain size variation but also to changes brought about by the metamorphism to which they have been subjected. The minerals present are quartz (60 to 90 percent), feldspar (plagioclase, microcline, and microperthite) (5 to 25 percent), and a fine matrix (5 to 20 percent) now largely sericite (with low-grade metamorphism) and biotite (higher grade of metamorphism). Metamorphism is least in the northern sequence and becomes progressively greater toward the southeast. In the unmetamorphosed rocks the quartz grains are rounded to subangular and occasionally show overgrowths; the grain contacts are of sedimentary type. Feldspar is fresh and in more angular grains than the quartz. A detrital matrix of fine grains, quartz and presumably feldspar, may be present to the extent of about 5 percent. Where metamorphosed, the original nature of these rocks is somewhat obscured by the alignment of quartz grains. In the early stages of metamorphism cataclastic breaking of the edges of quartz grains was common, and this resulted in an apparent increase in the detrital matrix of fine grains; sericite was developed as fine rods. As metamorphism intensified the sericite was wrapped around the quartz grains, which became much elongated with smooth edges, sometimes embayed. Strain effects were induced, and the original grain shapes were obliterated. The coarse feldspar was apparently more resistant to pressure than the quartz, for it often retains its original shape and size. At a higher grade of metamorphism biotite formed. Thus the rocks of the northern sequence are now schistose subarkoses, occasionally metaquartzites

with mosaic texture, whereas in the southern sequence the rocks are largely biotite metaquartzites often with strong microscopic schistosity.

Thick sequences of these rocks underlie most of the main range of the Great Smoky Mountains and many of the higher foothills to the north. In most places these arenaceous beds rest on older rocks of the Ocoee series with fault relationship. In the eastern part of the Great Smoky Mountains, however, the southern sequence lies directly on still older granite and gneiss with depositional contact (Goldsmith and Hadley, 1955). The southern sequence, which underlies the main range, is separated from that of the foothills by one or more thrust faults. Relations of the northern sequence are less clear than those of the southern sequence; faults separate it from other parts of the Ocoee series and from the southern sequence, but because the sense and magnitude of these faults is not known, the stratigraphic relationship remains obscure.

Saprolite, a term used by Becker (1895, p. 289) for "thoroughly decomposed, earthy, but untransported rock," is a common product of deep weathering of most of the rocks of this area, including the arenaceous beds of the Ocoee series. Textures and structures of the parent rock are preserved, but most of the weatherable minerals are altered. Profound leaching, oxidation, and hydration have removed such minerals as calcite and apatite; feldspars are altered to clay minerals; and pyrite and magnetite are oxidized. Many heavy minerals survive these processes thus making possible the use of saprolite for heavy mineral studies (Pardee and Park, 1948; Hurst, 1953; Mertie, 1954). By using saprolite the processing of the large samples required to obtain sufficient heavy minerals for examination is simplified as the material is already in a disaggregated state.

Fresh arenite and saprolite were compared mineralogically where a tongue of saprolite about 7 feet wide separates two exposures of fresh rock in a road cut (locality 24, fig. 1). Pebble layers are traceable across the saprolite between the fresh rocks, so that saprolite and fresh rock could be collected from the same bed. The following comparison was made by an examination of the two kinds of material:

Minerals in fresh rock	Appearance in saprolite
Quartz	Unchanged
Feldspar	Altered to clay mineral
Mica	Altered to hydromicas
Magnetite	Altered to hydrous iron oxides
Ilmenite	Altered wholly or in part to anatase
Apatite	Removed by solution
Zircon	Unchanged
Tourmaline	Unchanged

SAMPLES EXAMINED

Samples for the examination of the heavy minerals were collected from both the northern and southern arenite sequences to determine whether the mineral assemblages would show any differences between them. Evidence on the provenance of the detritus making up the arenites was also sought by

examining crystalline rocks from two sources: (1) granite pebbles from a conglomerate in the northern arenite sequence; and (2) crystalline rocks that underlie the Ocoee series. These crystalline rocks are exposed in uplifts that extend eastward from the area studied by the U. S. Geological Survey party (fig. 1). Richard Goldsmith, who has studied the uplift area, selected the sampling sites; the samples collected included coarse-grained granite from Max Patch Mountain and several kinds of gneiss.

The saprolite samples collected (see list at end of paper) fall into three main groups: samples 1-16B are a group of saprolite samples of arenite from scattered localities, including three pairs (3A and 3B; 4A and 4B; 16A and 16B) that were collected from the same or adjacent beds to check on local mineralogical variations; samples 25 to 29 are saprolites from a sequence of five beds in one exposure collected for further examination of bed-to-bed variation; and the remainder are samples of saprolite of crystalline rocks on which the Ocoee series rests. In addition, samples were collected to compare fresh rock with saprolite, to compare the mineralogy of rock and an included concretion, and to obtain the heavy minerals of granite boulders in a conglomerate in the northern arenite sequence.

METHOD OF EXAMINATION

Saprolite samples weighing 12 to 15 pounds were rolled to reduce the lumps and then sieved into grades of the Wentworth scale (1922) on a Ro-Tap. The fractions above 60 mesh (0.25 mm) were discarded, and the −60 +120 (0.125 mm) and the −120 +230 grades were quartered to reduce the bulk, and the remaining sample was then washed with water to remove the clay by agitation and decantation; then the samples were boiled for a short time in a 1:1 solution of HCl to remove iron coatings from the individual grains. The cleaned fractions of the smaller grade size (+230 mesh) were separated in bromoform to obtain the heavy minerals. No records were kept of the weight of heavy minerals yielded by these saprolite samples.

Samples of fresh granite (no. 30) and fresh arenite (no. 24B) were crushed and passed through rolls until all had passed the 80-mesh sieve (0.177 mm). The bulk of these samples was removed on a Wilfley table, thus obtaining a crude concentrate of the heavy minerals. A further concentration was made in bromoform, as for the saprolite samples. The heavy residues from the fresh rocks were weighed to give some indication of the amount of yield of heavy minerals. The granite sample weighed 6 pounds and contained 5 grams of heavy minerals, of which only 0.3 gram was zircon; the arenite sample weighed 2 pounds and yielded 6.25 grams of heavy minerals, of which 0.75 gram was zircon.

A portion of all heavy residues was mounted in Canada balsam, and the percentage mineral composition was obtained by grain counts.

MINERALOGY

The heavy residues of the saprolites had either a light or a dark appearance which was caused by a predominance of zircon or of ilmenite.

TABLE 1

Minerals Present in the Heavy Residues of Saprolite Samples of Arenaceous Beds of the Ocoee Series (percent by number from grain counts)

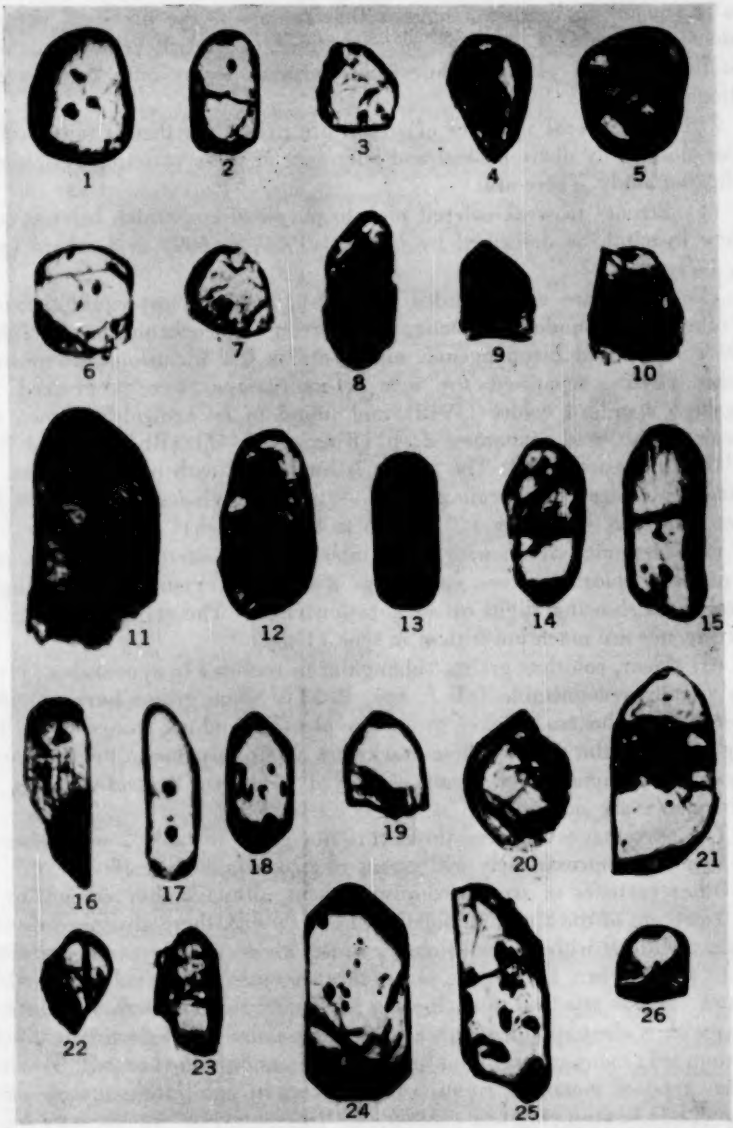
Sample no.	Ilmenite	Anatase	Zircon	Tourmaline	Others
			Northern sequence		
1	85	—	13	2	—
2	13	—	85	2	—
4A	44	—	54	2	—
4B	80	—	20	—	—
5	28	—	72	—	—
6	53	1/2	42	4	1/2
7	—	60	39	—	—
14	—	38	60	2	—
15	—	25	71	4	—
16	—	92	3	5	—
16A	—	71	15	14	—
			Southern sequence		
3A	2	—	98	—	—
3B	1	—	99	—	—
8	60	—	38	2	—
9	43	—	55	2	—
10	83	—	15	2	—
11	25	1/2	72	1 1/2	—
12	2	—	98	—	—
13	2	—	98	—	—
24B*	80	—	1	—	19 (apatite)
25	87	—	13	+	—
26	44	—	22	34	—
27	84	—	16	+	—
28	89	—	11	—	—
29	78	—	22	—	—

* Fresh arenite.
\+ One or two grains present.

The minerals identified, in order of abundance, are: zircon, ilmenite, anatase, tourmaline, rutile, and epidote. Zircon and ilmenite are present in varying amounts as shown in table 1 and together make up more than 90 percent of the total heavy minerals in the majority of the residues.

Ilmenite and anatase.—Ilmenite occurs as irregular composite grains encrusting or associated with micaceous grains, and seldom as well rounded individual grains such as are commonly found in sedimentary rocks or soils. Ilmenite with the same habit was observed in the thin sections of some of the arenites which show the effects of metamorphism. The encrusting habit gives the ilmenite a somewhat drusy appearance. In some residues the ilmenite is fresh, but in others there is an alteration in patches to leucoxene or to an encrusting brown to yellow titaniferous mineral, either anatase or rutile, indeterminate because of its small size. There is almost complete replacement of ilmenite by anatase in five of the samples (nos. 7, 14, 15, 16, 16A; table 1). Authigenic anatase in pale steel-gray octahedra of typical habit is found in

PLATE 1



Zircon grains in arenaceous beds of the Ocoee series and in the Max Patch granite. (Magnification approx. ×100.)

Figs. 1-7, purple or hycinth zircon, Type (1).

Figs. 8-13, metamict zircon, Type (2).

Figs. 14-21, colorless zircon, Type (3).

Figs. 22-26, Zircon from the Max Patch granite, sample 17A.

(Grains were mounted in oil with refractive index of 1.73)

a number of the residues, and this, together with recognizable anatase frequently encrusting ilmenite, suggests that anatase is the principal alteration product of ilmenite. Leucoxene, here confined to whitish yellow opaque encrustations, is not plentiful, but where present forms only thin irregular coatings.

Zircon.—Several varieties of zircon are present in these residues. Three of the most easily distinguished and abundant of these varieties were selected for further study. These are:

(1) Strong- to weak-colored pink to purple zircon which belongs to the variety hyacinth as described by Hutton (1950, p. 689) and others (pl. 1, figs. 1-7).

The grains are well rounded and fresh, many almost equidimensional, but others are rounded prismatic. This zircon is pleochroic, has a high refractive index and birefringence, and contains few inclusions. Concentrates of this zircon, separated for age determinations, were compared with Ridgway's standard colors (1912) and found to be brownish purple, deep brownish drab, and cinnamon drab (Ridgway 5′′′00.Rb, I′′′Red B, or 5′′′00.Rd, respectively). The purple is intensified with polarized light, and in strongly colored pleochroic grains ω = purplish vinaceous, and ε = light russet vinaceous (Ridgway I′′′Red b to 9′′′OR-Ob).

(2) Metamict zircon with dense interiors often surrounded by an outer rim of clear, colorless zircon (pl. 1, figs. 8-12), well crystallized with angular crystal faces showing slight or no abrasion effects. The refractive index and birefringence are much lower than in type (1).

(3) Clear, colorless grains, subangular to rounded in appearance; crystal faces usually recognizable (pl. 1, figs. 13-21). Some grains have abundant, rather large inclusions. Broken grains are plentiful and are caused by splitting along cracks in the grains. These cracks are apparently due to the presence of elongated inclusions which create a zone of weakness. Refractive index and birefringence are normal.

The percentages of these three varieties given in table 2 were obtained by a count of approximately 100 grains of zircon in each residue.

Other varieties of zircon are also present, although they amount to less than 1 percent of the zircon population. For example, there are zoned crystals, colorless, with or without inclusions; grains with very flat pyramids, probably (101), which when broken appear square in cross section; a few doublets, i.e., two crystals attached to each other parallel to the *c* crystallographic axis; grains with a clear rounded central core surrounded by clear zircon which is also rounded; many cracked and broken grains as described under (3) above; another type of metamict zircon which differs in habit from that described under (2) in having very fine zoning.

Tourmaline.—Tourmaline is for the most part in angular broken fragments although in one or two of the residues there are some slightly rounded grains. There is little diversity in color of this tourmaline; most of the grains are either brown or gray. In nos. 14 to 16A tourmaline is slightly more abundant than in the other residues, and in no. 26 it makes up 34 percent of the heavy minerals (table 1). No recrystallized tourmaline was recognized,

TABLE 2

Distribution of the Three Principal Varieties of Zircon in the Heavy Residues from Saprolites of Arenites of the Ocoee Series (percent by number in the zircon population)

Sample no.	Varieties of Zircon (1) Purple (hyacinth), fresh, mostly rounded	(2) Metamict, mostly angular	(3) Colorless, subrounded to rounded
	Northern sequence		
1	57	12	31
2	55	9	35
4A	31	12	57
4B	61	5	34
5	58	5	37
6	44	7	49
7	37	5	58
14	61	8	31
15	58	7	35
16	52	8	40
16A	51	5	44
	Southern sequence		
3A	5	7	88
3B	36	8	56
8	63	10	27
9	51	8	41
10	61	12	27
11	64	6	30
12	50	12	38
13	60	11	29
24A	41	8	49
24B*	29	7	63
25	40	11	49
26	29	—	71
27	40	8	52
28	37	10	53
29	42	19	39

* Fresh arenite

but the fragmentation may have been caused by the metamorphism to which the arenaceous beds have been subjected.

Other minerals.—A few grains of a colorless to very pale green mineral, altered and corroded, were observed in some of the residues. This is probably epidote. No metamorphic minerals such as garnet, staurolite, or the aluminous silicates were found. Monazite was absent.

To sum up the mineralogical composition of the heavy residues, it is apparent that:

(1) The arenaceous beds contain a very restricted suite of minerals.

(2) These minerals occur in a wide range of proportions.

(3) The proportions vary between samples.

(4) The northern and southern sequences of arenaceous beds are not characterized by distinctive mineral suites.

The heavy minerals are those commonly associated with granitic and gneissic rocks. Detrital minerals derived from metamorphic rocks are absent, as are ferromagnesian minerals from basic igneous rocks. Products of the regional metamorphism to which the rocks have been subjected are sericite and biotite, although elsewhere in the area garnet, staurolite, and kyanite are developed (Hadley and others, 1955, p. 409).

Ratios of zircon to ilmenite vary widely, as do the proportions of the three principal varieties of zircon (tables 1 and 2). Bed-to-bed variation in mineralogy was investigated at one exposure (nos. 25 tc 29) where a succession of several beds of saprolitized arenites was sampled. In this succession one bed (no. 26, 10 inches thick) has a heavy residue which is unlike that of the bed above (no. 25) or those beneath (nos. 27, 28, and 29), the main difference being in the very large quantity of tourmaline present (34 percent). The low proportion of hyacinth to colorless zircon in this bed is also quite noteworthy.

TABLE 3

Distribution of the Three Principal Varieties of Zircon in the Zircon Population of the Heavy Residues of Saprolitic Arenites of the Ocoee Series (percent by number)

Heavy residue contains:	High Zircon, 70 to 100%	Medium Zircon, 25 to 70%	Low Zircon 0 to 25%
	Percent	Percent	Percent
(1) *Purple or hyacinth*			
Range	5 to 64	31 to 63	29 to 61
Average	48	48	47
Median	56	47	46
Q_1	42	37	40
Q_3	59	61	57
(2) *Metamict*			
Range	5 to 12	5 to 12	5 to 19
Average	8	8	9
Median	7	8	9
Q_1	6	7	5
Q_3	10	10	12
(3) *Colorless*			
Range	29 to 88	27 to 58	27 to 71
Average	43	48	44
Median	36	53	42
Q_1	32.5	41	34
Q_3	47	58	52

Q_1, percentage at the 25 percent point in the distribution.
Q_3, percentage at the 75 percent point in the distribution.

Proportions of the three principal varieties of zircon appear to be independent of the total amount of zircon present (table 3). These variations are to be seen not only in the more widely spaced samples but also in adjacent beds (nos. 25 to 29) and in different parts of the same beds (nos. 3A and 3B; 4A and 4B). Closer examination might reveal that the more distinctive assemblages characterize certain beds, as was found to be true in the Ocoee

series in the Ducktown district (Gibson, 1953), but this possibility could not be explored in this study because of the discontinuity of outcrops and resulting uncertain correlations.

TABLE 4

Distribution of the Three Principal Varieties of Zircon in the Heavy Residues of Saprolite of the Northern and Southern Arenite Sequences (percent in the heavy residue from grain counts)

	Northern sequence	Southern sequence
	Percent	Percent
(1) *Purple or hyacinth*		
Range	31 to 61	5 to 64
Average	51.3	44.5
Median	55	42
Q_1	44	36.5
Q_3	58	60.5
(2) *Metamict*		
Range	5 to 12	6 to 19
Average	7.5	9.3
Median	7	10
Q_1	5	7.5
Q_3	9	11.5
(3) *Colorless*		
Range	31 to 58	27 to 88
Average	41	46.2
Median	37	41
Q_1	34	29.5
Q_3	49	54.5

Q_1, percentage at the 25 percent point in the distribution.
Q_3, percentage at the 75 percent point in the distribution.

Comparison of mineral assemblages from northern and southern arenaceous beds shows complete overlap, or differences so small as to be negligible (table 4). The field criteria that permitted distinction between the northern and southern sequences of beds is thus not reflected in the heavy minerals.

MINERALOGY OF CRYSTALLINE ROCKS

Granite.—Boulders of gray, coarse-grained granite from a boulder bed in the northern arenite sequence provided an opportunity to examine at least a part of the presumed source of the detrital material (sample 30). In the 5 grams of heavy residue obtained from 6 pounds of this granite are pyrite, 0.11 percent (percentages by weight of the whole rock); magnetite and ilmenite, 0.04 percent; zircon, 0.01 percent; others (epidote, monazite, sphene, tourmaline), 0.02 percent. The dominant variety of zircon is brownish purple, zoned and metamict; it is in slightly rounded prismatic grains with no recognizable pyramidal faces. Many of these zircon grains have a pronounced brown central core of opaque material; less than 10 percent of the grains are fresh

and colorless. The brownish-purple zircon of this granite was not recognized in the residue from any arenite saprolite sample.

Crystalline rocks below the Ocoee series.—Crystalline basement rocks underlie the Ocoee series in uplifts in the southeastern part of the Great Smoky Mountains (fig. 1) (Goldsmith and Hadley, 1955). This basement consists of a gneiss complex (Carolina gneiss) intruded by the Max Patch granite. One sample of saprolite of the Max Patch granite (no. 17A) and three samples of the gneiss complex saprolite (nos. 18, 19B, and 20B) were collected in order to compare their heavy residues with those of the arenites, particular attention being paid to the zircon varieties.

The saprolite of the Max Patch granite contains epidote and zircon as stable heavy minerals. The zircon concentrate has a slightly brownish-pink color in mass (Ridgway, light cinnamon drab). The individual grains are prismatic with rather squat simple pyramids (pl. 1, figs. 22 to 26). Many are badly cracked but most contain relatively few inclusions, and these tend to be rounded rather than elongated. Some grains are dense from innumerable internal cracks which could in part be produced by late low-temperature post-Ocoee deformation. Most of the crystal faces are sharp edged, and many grains are broken; zoning occurs but is not common. No metamict zircon was seen. To judge from this sample, therefore, the zircon grains from the Max Patch granite are not similar to any of the important varieties in the arenaceous beds.

The zircon grains in the gneiss saprolite samples are colorless, cracked, and occasionally zoned. They are accompanied by epidote and possibly a little monazite (in sample 19B). Sample 20B contains many zircon grains with longitudinal striations and very few inclusions; many of these are rounded and have no fresh crystal faces. Many surfaces are marked by irregular cracks. No metamict zircon was found in these samples. The zircon in these gneiss samples does not appear to be comparable with those in the arenites.

AGE DETERMINATIONS OF ZIRCON

Age determinations were made of 9 fractions of zircon separated from 6 samples of saprolitic arenite and of 1 fraction from saprolite of the Max Patch granite. The ages were determined by the total lead-alpha activity ratios of the zircon according to the method of Larsen, Keevil, and Harrison (1952), where $T_1 = \frac{Pb \times 2400}{alphas/mg/hr}$ for samples whose age is less than 300 million years. For older samples, the parent decay of uranium requires a correction of T_1 according to the formula: $T = T_1 - 1/2(T_1^2 \times 1.55 \times 10^{-4})$. The correction increases with time (Keevil, 1939) and for zircon of Late Precambrian age is in the order of 6 percent. In this method the lead is determined spectrographically and is reproducible to 6 to 10 percent (Waring and Worthing, 1953). The alpha counts are made on 100 mg of finely pulverized zircon in an ionization chamber receiving a continuous flow of argon. Alpha counts are reproducible to 5 percent. Limits of error of the age determination

method have not been determined. Reproducibility of determinations has been ± 10 percent in previous studies.

In the heavy residues more than one variety of zircon was found in each sample of arenite saprolite examined. It was found possible to separate fresh from metamict zircon by varying the pitch, tilt, and amperage of the Frantz isodynamic separator, as zoned and metamict zircon is commonly slightly more magnetic than fresh, clear zircon. For three of the samples (nos. 2, 9, and 12) fractions of both fresh and metamict zircon could be separated. In each of the three, different ages were obtained for the two different zircon fractions in a single sample of saprolite. Three additional samples of arenite saprolite (nos. 3B, 8, and 13) did not yield a sufficient amount of metamict zircon for an age determination. The zircon separates used for age determination were as nearly homogeneous as could be obtained and were approximately 90 to 95 percent of one variety after magnetic separation and hand picking.

The results of the age determinations are given in table 5 where samples are grouped with respect to variety of zircon and numerical age. Zircon samples of similar morphology give similar ages (table 5). Rounded, fresh, purple zircon found in samples 2, 8, and 9 gives ages of 1030, 1140, and 1120 million years, respectively, with an average age of 1097 ± 6 percent. Metamict, well terminated zircon found in the same rocks (nos. 2 and 9) gives ages of 830 and 870 million years. Sample no. 8 did not contain enough metamict zircon for an age determination. Elongated, fresh, purple zircon found in samples no. 12 and no. 13 both give ages of 620 million years. This grouping of results suggests that the Ocoee arenites may have received zircon from three sources in Precambrian time at approximately 600, 850, and 1100 million years. Single age determinations on two additional morphological

TABLE 5

Lead-alpha Activity Age Determinations of Zircon from Six Samples of Saprolitic Arenites of the Ocoee Series and of One Sample from Saprolite of the Max Patch Granite, Great Smoky Mountains, Tennessee and North Carolina.

Zircon no.	Sample no.	Alphas/mg/hr*	Pb (ppm)**	Age (million years)	Variety of zircon
12A	12	129	35	620	Fresh, purple; elongated, some grains rounded.
13A	13	136	37	620	" " "
12M	12	300	99	740	Metamict, moderately rounded.
3B	3B	153	61	890	Fresh, colorless, rounded.
2M	2	287	107	830	Metamict, well terminated.
9M	9	273	107	870	" " "
2A	2	124	58	1030	Fresh, purple, rounded.
9A	9	112	58	1120	" " "
8A	8	110	58	1140	" " "
17A†	17A	78	31	880	Fresh, plae brown, well terminated.

* Determined by H. W. Jaffe, U. S. Geological Survey.
** Determined by C. L. Waring, U. S. Geological Survey.
† 17A—Saprolite of the Max Patch granite.

varieties of zircon give ages of 740 and 890 million years, and additional determinations may warrant their assignment to the age grouping averaging 850 million years.

The zircon of intrusive rocks is frequently not of one type or variety, even in the same specimen. Clear fresh zircon may occur with highly metamict subtranslucent zoned zircon in the same sample of intrusive rock. Nevertheless the associated zircon crystals, however different in apperance, uranium and lead contents, will all give the same age to ± 10 percent. It is obvious that zircon from several sources may be present in any sample of sedimentary rock, and petrographers have made use of the varietal features of zircon grains to trace the probable source of heavy detrital residues (Brammall, 1928; Groves, 1931; Poldervaart, 1955).

The final "homogenous" fractions on which these ages were determined may in reality be mixtures derived from more than one source. The age determinations of the zircon, therefore, do not give the age of the arenites but merely a range of time for the crystallization of the rocks from which the detrital grains were derived. Although the youngest age determined for a zircon separate was 620 million years, it does not prove that the arenites themselves are Late Precambrian.

SUMMARY OF INFORMATION ON HEAVY MINERALS

The heavy minerals in the saprolitic arenites consist of ilmenite and zircon, with sporadic enrichments of tourmaline. Ilmenite is a common detrital mineral in many kinds of sedimentary rocks and therefore has little significance for determining provenance. Its irregular grain shape and association with micaceous grains is due to the metamorphic processes which have affected these rocks. Ilmenite has altered to anatase in the weathering processes whereby the arenites were changed to saprolites.

Zircon is characteristic of granite and gneiss, and the varietal types are of interest and importance in suggesting parentage. The average composition of the zircon population in 26 samples of arenite saprolite is: hyacinth or purple, 48 percent; metamict, 8 percent; colorless, 44 percent. The ranges in percentages are given in tables 3 and 4, from which it can be seen that there is no significant difference in distribution between the northern and southern sequences. The ages of the zircon range from 1140 to 620 million years (table 5).

Saprolites of gneiss and granite beneath the Ocoee series contain varieties of zircon different from those in the arenites.

In the fresh arenite and in the only concretion examined, magnetite and apatite accompany ilmenite, zircon, and tourmaline. There was little difference in quantity of minerals between the fresh rock and the concretion. Apatite and magnetite are removed by the formation of saprolite.

DISCUSSION OF RESULTS

Interest in the heavy residues, and particularly in the zircon content of the arenites and of the granite and gneissic rocks beneath, centers on the use which may be made of them as indicators of provenance. About half of the

zircon is a deeply colored, purple variety similar to that described from rocks of Precambrian age in many parts of the world; for example, Scotland (Mackie, 1923), Canada (Tyler and others, 1940), New Zealand (Hutton, 1950), Australia (Carroll, 1939). The ages of the zircon obtained from the Ocoee series suggest that they were derived from a source in Precambrian rocks.

The presence of different varieties of zircon and the range of their ages suggest that these arenites of the Ocoee series received detritus from several source rocks. Variations in the proportions of the minerals in the heavy residues appear to be due to local variations in deposition rather than to major changes in source of material. Samples 25 to 29 (table 1) provide an example of such variation.

Because of the greater degree of rounding of the purple zircon (type no. 1) in comparison with other types, together with its greater age, it is thought that there may have been two principal sources of the detrital minerals; one source was probably a Precambrian gneiss similar in mineralogy to the Scottish Lewisian gneiss which supplied the purple zircon to many of the sedimentary rocks of Britain (Mackie, 1923). Another source may have been a granite or a granite-gneiss complex which supplied the colorless more angular zircon and the metamict zircon together with tourmaline. Metamict zircon may be regarded as a less stable form of zircon from the fact that it is not commonly present in the residues of sandstones whose detrital grains have been through several cycles of sedimentation. The angular prismatic grains of metamict zircon strongly suggest that it came directly from a granite, possibly along with the colorless zircon. The ages of these two varieties of zircon have been determined to be approximately the same.

The tourmaline grains in some of the heavy residues are well rounded but in others they are angular. In the least metamorphosed rocks examined, those with rounded quartz grains, a few well rounded grains of tourmaline were observed. The fragmentary angular character of most of the tourmaline in the heavy residues need not necessarily imply a nearness to source nor a lack of abrasion, for the metamorphic processes which destroyed the original shapes of the quartz grains could have shattered the tourmaline grains as well. On the other hand, zircon grains observed even in the more strongly metamorphosed sediments preserve their original shapes.

Another possibility as a source for the heavy minerals could be some Late Precambrian sediment which contained all three principal varieties of zircon in about the proportions that they occur in the arenites.

It is pertinent here to note further points made by Mackie (1923) as a result of his investigations of Scottish granites. When he compared the content of purple zircons in gneiss and in grit (sandstone with angular grains) derived from it, he found that the quantity of these zircons indicated that there had been very little material lost in the change from gneiss to grit. He noted too, that the granite which intruded the Lewisian gneiss picked up purple zircon from the gneiss and added it to the types peculiar to its own magma. The types of zircon which Mackie described from granites include what are now known as metamict zircon. Mackie found that the rocks richest

in purple zircon were grits which were obviously derived from the Lewisian gneiss.

Two other examples of sedimentary rocks somewhat similar to the arenites in the Ocoee series have been described recently: one from South Africa (Poldervaart, 1955) and the other from Australia (Clarke, Phillipps, and Prider, 1954).

In the arenaceous sediments of the Witteberg, Bokkeveld, and Table Mountain series, Cape Province, described by Poldervaart (1955), only five nonopaque heavy minerals are present: zircon, rutile, brookite, tourmaline, and monazite. In 17 samples described (Poldervaart, 1955, p. 451) the percentages are: zircon, 54 to 81 (average, 65); rutile, 7 to 25 (average, 19); brookite, trace to 16 (average, 6); tourmaline, 2 to 15 (average, 6); and monazite, trace to 4 (average, 2). No details of the proportion of opaque to nonopaque heavy minerals in these rocks were given, so that no further comparisons can be made. Although the source of the zircon has not definitely been traced, it is suggested from the presence of elongated zircon with steep pyramidal faces, known to occur in the Cape granite, that this granite was one of the sources of these rocks, particularly of the Witteberg and Bokkeveld series which are at the base of the arenites.

Mica schists at a low grade of metamorphism, similar to that of some of the rocks of the Ocoee series described in this paper, have been described from Western Australia by Clarke and others (1954) as containing varying proportions of ilmenite and zircon (including a rounded purple variety) together with minor amounts of anatase and tourmaline. The anatase has developed authigenically from the alteration of ilmenite. Like the arenites in the Ocoee series, some beds of this mica schist contain ilmenite as the predominating heavy mineral (ilmenite, 60 percent; zircon, 20 percent). These rocks are considered to be of very Late Precambrian age.

The source of the purple zircon in the arenites of the Ocoee series is not known, and until further investigations are made of the varieties of zircon in the gneisses and granites of the southern Appalachian region, the suggestions made here are merely speculative. Neither the Max Patch granite (pl. 1, figs. 22-26) nor the gray granite collected from a boulder bed in the Ocoee series contains zircon similar to that in the arenites nor do several gneisses below the Ocoee series.

CONCLUSIONS

The examination of heavy residues of a number of samples from two sequences in the Ocoee series which were mapped in the field as separate units has failed to reveal any mineralogical differences of significance (tables 3 and 4). On the contrary, close similarity of the heavy-mineral assemblages from both northern and southern sequences indicates that they had a common source. Local variations in deposition could account for the varying proportions of the two principal heavy minerals, zircon and ilmenite, in individual beds.

The mineral assemblage as a whole does not contain minerals which are exclusively common to granites, but on the other hand it contains no min-

erals, except ilmenite, which are common to basic igneous rocks. The simplicity of this assemblage together with the rounded form of quartz and tourmaline grains in the least metamorphosed parts of the arenites suggest the possibility of some pre-Ocoee sorting between the original source or sources of the zircon and tourmaline and its deposition as a sediment. However, processes producing disintegration of the immediate source rocks were apparently sufficiently rapid to prevent the removal of apatite and the formation of clay; or if clays did develop, then they were removed prior to deposition of the arenites. Preservation of detrital feldspar and apatite attest to the special conditions of weathering and erosion that prevented their destruction —governed either by arid climate under which chemical weathering is at a minimum or by relief which is sufficiently great to promote denudation over chemical weathering. The poor sorting and graded bedding that characterize these arenites suggest the likelihood of the latter.

The paucity of mineral species is more difficult to explain satisfactorily. The source rocks themselves may have had an equally limited suite, but Pettijohn (1941) has remarked on similar lack of heavy minerals in other older rocks and suggested that intrastratal solutions were responsible for the removal of less stable minerals. Bramlette (1941) demonstrated the effectiveness of intrastratal solutions by showing that mineral species were preserved in concretions but not in the surrounding rock. The concretion examined in this investigation failed, however, to reveal significant differences. It seems probably that predepositional sorting is responsible for the removal of some heavy minerals from a source that was originally poor in the number of species which it had to offer.

The minerals which are found in the heavy residues of these arenites suggest a granitic and gneissic provenance, but a comparison between the assemblages of heavy minerals from the arenites of the Ocoee series and from the local gneisses has so far failed to reveal the source of the minerals. The ages determined for several varieties of zircon indicate that rocks of more than one source contributed zircon to the sediments.

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LIST OF SAMPLES EXAMINED

1. Saprolite from arenite, 2.05 mi N 80° W of Gatlinburg (school), Gatlinburg quad., Tenn.
2. Saprolite from arenite, 1.95 mi N 85° W of Gatlinburg (school), Gatlinburg quad., Tenn.

3A. Saprolite from conglomeratic arenite, U. S. Highway 441, opposite mouth of Ramp Creek, Gatlinburg quad., Tenn.

3B. Saprolite from conglomeratic arenite, U. S. Highway 441, opposite mouth of Ramp Creek, Gatlinburg quad., Tenn.

4A. Saprolite from arenite, Laurel Falls trail, 3/4 mi W of Fighting Creek Gap, Gatlinburg quad., Tenn.

4B. Saprolite from arenite, Laurel Falls trail, 3/4 mi W of Fighting Creek Gap, Gatlinburg quad., Tenn.

5. Saprolite from arenite, U. S. Highway 129, 1.11 mi E of Calderwood P. O., Calderwood quad. Tenn., N. C.

6. Saprolite from arenite, U. S. Highway 129, 1.15 mi E of Calderwood P. O., Calderwood quad., Tenn., N. C.

7. Saprolite from arenite, U. S. Highway 129, 1.50 mi E of Calderwood P. O., Calderwood quad., Tenn., N. C.

8. Saprolite from arenite, U. S. Highway 129, 1.80 mi E of Calderwood P. O., Calderwood quad., Tenn., N. C.

9. Saprolite from arenite, U. S. Highway 129, 2.09 mi E of Calderwood P. O., Calderwood quad., Tenn., N. C.

10. Saprolite from arenite, U. S. Highway 129, 2.5 mi S 80° E of Calderwood P. O., Calderwood quad., Tenn., N. C.

11. Saprolite from arenite, U. S. Highway 129, 1.94 mi N 30° W of Deals Gap Tapoco quad., Tenn., N. C.

12. Saprolite from arenite, Bote Mountain road, alt. 2500 ft., Thunderhead quad., Tenn., N. C.

13. Saprolite from arenite, Bote Mountain road, alt. 3000 ft., Thunderhead quad., Tenn., N. C.

14. Saprolite from arenite, old Cades Cove road, 0.5 mi N 50° E from Cold Spring Gap, Kinzel Springs quad., Tenn.

15. Saprolite from arenite, old Cades Cove road, Cold Spring Gap, Kinzel Springs quad., Tenn.

16A. Saprolite from arenite, Old Cades Cove road, 0.5 mi N 15° E from old C.C.C. camp, Cades Cove quad., Tenn., N. C.

16B. Saprolite from arenite, locality as for 16A.

17A. Saprolite from granite, about 1 mi N of Max Patch Mountain on Lemon Gap road, Lemon Gap quad., N. C.

18. Saprolite from gneiss, 2 mi N N E of Fine's Creek school, Hepco quad., N. C.

19B. Saprolite from gneiss, 0.3 mi E of Fine's Creek school, Hepco quad., N. C.

20B. Saprolite from gneiss, North Carolina State Highway quarry, 1 mi W of Crabtree on State highway 209, Clyde quad., N. C.

23A. Concretion in arenite, U. S. Highway 129, N. of Cooper Camp Branch, 0.6 mi W N W of BM. K11, Tapoco quad., Tenn.

23B. Matrix of concretion, locality as for 23A.

24A. Saprolite from arenite, locality as for 23A.

24B. Fresh arenite, locality as for 23A.

25 through 29. Samples from various beds of saprolitic arenite, 0.85 mi S W of Cold Spring Knob, U. S. Highway 129, Calderwood quad., Tenn.

30. Granite boulders in conglomeratic bed in arenite, headwaters of Davis Branch, Kinzel Springs quad., Tenn.

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ON THE MECHANISM BY WHICH STONES IN TILL BECOME ORIENTED

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ABSTRACT. In most till deposits the long axes of the stones tend to be aligned parallel to the direction in which the ice flowed. Sometimes there is also a tendency for long axes to be oriented transverse to the flow direction. This transverse peak is particularly noticeable in narrow band tills. Laboratory experiments and theoretical studies by various authors have shown that an initially random collection of elongated objects immersed in a flowing liquid will develop a long axis distribution with a peak parallel to the direction of flow in a very short time, but that if flow is continued for a long time a transverse peak will also develop. The observed orientations could thus be entirely the result of flow in moving ice. Collisions between particles and the mode of deposition could change the relative distribution of long axes between the two peaks. The theoretical results are compared with distributions of long axes found in various tills.

INTRODUCTION

The purpose of this paper is to discuss the mechanisms by which stones in glacial deposits can acquire a preferred orientation with respect to the direction of flow of the ice which deposited them. In order to refer to the various directions relative to the flow of ice, let us consider a block of ice forming part of the bottom of a glacier flowing over a plane bed as shown in figure 1. The direction parallel to the direction of flow will be referred to as the *parallel* direction, the direction perpendicular to the direction of flow but parallel to the bed will be referred to as the *transverse* direction, and the direction perpendicular to both the direction of flow and the bed will be referred to as the *perpendicular* direction. The plane containing the parallel and perpendicular directions will be termed the *plane of shear*, the plane containing the parallel and transverse direction will be termed the *parallel plane*, the plane perpendicular to the direction of flow will be termed the *perpendicular plane*. All of these terms are shown in figure 1.

It has long been known that the stones or rock fragments in till are not oriented at random, but that the long axes of stones tend to have a preferred orientation in the parallel direction. It is sometimes also found that a subsidiary peak of long axis orientations lies in the transverse direction. Holmes

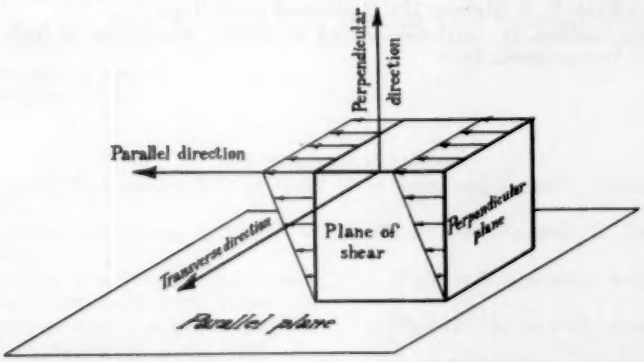

Fig. 1. Diagram to show the terms used for the different planes and directions in a material flowing by simple shear.

(1941) made a detailed study of the preferred orientations of stones in till in central New York State, determining not only the long axis directions but also the directions of the intermediate axes, the form and the roundness of each of over 1000 stones. He noted various systematic tendencies between various characteristics of the stones, and also discussed the mode of formation of the peaks in long axis distribution. He attributed the prominent parallel peak to orientation by gliding of stones in contact with the glacier floor (or possibly along well-defined shear planes in the ice), while he attributed the transverse peak to the tendency of stones totally immersed in the ice to assume an orientation in which they rotate about their long axes.

If these were the only mechanisms which could produce preferred orientations of stones, then conclusions about the nature of ice flow could be deduced from observations of the number of stones near the two peaks in the long axis distributions. Thus a till built up beneath active ice would be expected to give a strong parallel peak, while a till formed by deposition from dead ice might be expected to show a more prominent transverse peak, inherited from the orientations assumed in the flowing ice. However there are physical reasons for suggesting that the parallel orientation can also be produced in shearing ice, without the necessity for any dragging or sliding. This being so, it seems opportune to review all the possible methods by which preferred orientations may have developed, to see what deductions, if any, can be drawn from observations of stone orientations. These results can be compared with measurements of stone orientations taken on tills still surrounded by ice in Nordaustlandet (Donner and West, 1956), where large transverse orientations appeared to be formed only in narrow band tills, and with orientation measurements on deposits in eastern England (West and Donner, 1956).

Although the considerations brought forward in this paper are related to the orientation of stones, similar considerations will apply to the orientation of the microscopic particles in till, although the importance of the various factors will be rather different. Seifert (1954) has studied the preferred orientations in this case, and his results also can be considered in the light of the physical arguments here suggested.

POSSIBLE CAUSES FOR PREFERRED ORIENTATION OF STONES IN TILL

The causes which could be responsible for preferred orientations can be divided into three main groups, those stemming from the method of entrainment of the stones in the ice, those inherent in the flow process itself, and those resulting from deposition or subsequent movement of the deposit.

Entrainment of debris in the ice.—The debris which forms the stones in till may enter the ice in several ways. It may be plucked from rocks which are being shattered by freezing, it may be washed along a melt water stream, or it may fall onto the surface of the glacier from rock walls. All of these processes will place stones in the ice in initial orientations that are probably not random. For example, stones are unlikely to enter the ice with their long axes vertical. It is difficult, however, to estimate exactly what distribution is to be expected, and probably the best assumption to make in computing the effects produced in the subsequent stages is to assume that the stones do in

fact enter the ice quite randomly oriented. This assumption is unlikely to lead to spurious preferred orientations.

Flow in the ice.—A stone completely immersed in ice which is flowing by shear will be rotated by the ice as well as being carried forward. The rotation rate depends on the orientation of the stone, and does not remain constant as the stone is moved. The theory of the movement of non-spherical objects suspended in a flowing liquid has been studied for the case of ellipsoids in a viscous liquid by Jeffery (1922). His calculations show that, provided the velocity of flow is small, the ellipsoids, as well as being carried forward, rotate in orbits about their centers, their long axes returning after a time to their original orientation. The rate of rotation varies, being least when the long axis of the stone is most nearly parallel to the direction of flow. For example, for the simple case of a prolate spheroid (i.e. an object shaped like a rugby football) whose long axis is initially in the plane of shear (fig. 2), the long

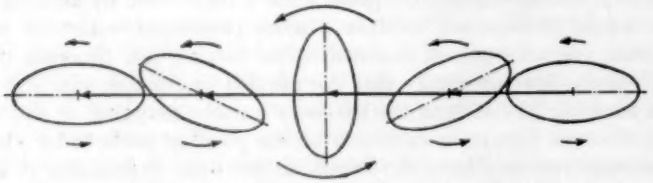

Fig. 2. Diagram to demonstrate the movement and rotation of an ellipsoidal object in a shearing material.

axis rotates so that it remains in this plane, but the angular velocity has a maximum when the long axis is in the perpendicular direction (the position in the center of fig. 2) and a minimum when it is in the parallel direction (at the left and right in fig. 2). The ratio of the maximum to the minimum angular velocity is the square of the axial ratio of the spheroid. The spheroid therefore spends much more time with its long axis near to the parallel direction than it does perpendicular to this. This result is very reasonable, as, when the long axis is perpendicular the top and bottom of the spheroid extend into regions of the fluid flowing relative to the centre of the spheroid at a faster rate than the regions reached when the long axis is parallel in direction.

For a prolate spheroid in a more general orientation the result is similar but more complicated. Jeffery's calculation shows that the angular velocity of the long axis about an axis in the transverse direction still has a minimum when the long axis is in the parallel plane, but there is a further effect, for the angle between the long axis and the plane of shear also varies cyclically and is least when the long axis lies in the parallel plane, and greatest when the axis is in the perpendicular plane. This effect crowds the long axes together towards the parallel direction, and so increases the effect due to the angular velocity variations by which, at any given time, there is a greater chance of finding a given spheroid near to the parallel direction. The preferred orientation resulting from this flow can be calculated from Jeffery's formulae if the initial positions of the spheroids are known. A completely random distribution of spheroids is not stable, i.e. if it existed at one time, then a short time later

the flow would have produced a preferred orientation. If it is assumed that the stones are placed in the liquid in random orientations, then the number of stones in any given orbit will be proportional to the angle subtended by that orbit at the centre of the spheroid, i.e. to the number of different orientations that lead to that orbit. However within any one orbit, the chance of finding a spheroid in any given part of the orbit will be inversely proportional to the angular velocity at that part. In this way the probability of any given long axis orientation can be deduced, and from this a rose diagram can be constructed to show the distribution of the projection of long axes on the parallel plane. This computation has been carried out graphically for the case of an axial ratio of 2:1, and the resulting rose diagram is shown in figure 3. This

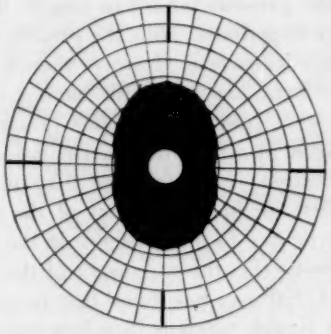

Fig. 3. Long-axis rose diagram computed from Jeffrey's theory for prolate spheroids with an axial ratio of 2:1 in a linear viscous liquid assuming the spheroids initially to have been placed in the liquid at random. The top and bottom of the diagram represent the orientation parallel to the direction of flow.

diagram has been drawn so as to be directly comparable with the rose diagrams constructed from measurements on tills, the radii vectores are measured from the circumference of the small white center circle, and are proportional to the number of spheroids which would have their long axes in the plane perpendicular to the parallel plane and containing the direction of the radius vector. The parallel direction is at the top and bottom of the diagram, the transverse direction at the sides. It will be seen that for this axial ratio, there are just over twice as many stones lying near the parallel direction as near the transverse direction.

In the solution of the viscous problem given by Jeffery there is no tendency of the spheroids to move from orbit to orbit. Thus after all the spheroids have rotated once or twice, the preferred orientation pattern should not change with further flow. In the transverse direction there is a minimum in the rose diagram, and there is no obvious reason why this situation should change. However Jeffery himself pointed out that the energy dissipated by the spheroid is a minimum if the long axis is in the transverse direction. A similar result is true for a more general ellipsoid, while for an oblate spheroid the minimum energy is dissipated if it rotates about a diameter (in this case the discus-shaped object has no definite longest axis, and a diameter is just one of the

infinite number of longest axes). Jeffery suggested that higher order terms in the equation, which he had neglected, might be expected to favor the orientation which dissipates minimum energy, though it is to be expected that such a reorientation might take a very long time, and as the viscosity enters into the terms neglected, the higher the viscosity, the longer one might expect this time to be.

Various experiments have been performed to check the results of Jeffery's calculation and his surmise about the transverse orientation. Taylor (1923) immersed both prolate and oblate spheroids in water-glass and studied their behavior during shear between concentric cylinders. He found that the ellipsoids did rotate in orbits of the kind predicted by Jeffery, and he also found that after a long time the particles tended to assume the orientation predicted by the minimum energy hypothesis. Thus the prolate spheroids slowly shifted until their long axes were transverse. However the time taken to establish this reorientation was quite large, the spheroids having rotated about 370 times before their long axes were transverse. Taylor remarks how difficult it is to understand the mechanism by which this reorientation takes place, and Saffman (private communication) has suggested that the cause may not be the terms neglected by Jeffery in his calculation of the solution for the perfect viscous liquid, but may be due to the existence in the liquid of thixotropy, in particular of cross viscosity (i.e. the expansion of the liquid in the transverse direction when sheared). Saffman has found that reorientation occurred quite rapidly in a thixotropic liquid, whereas in a Newtonian viscous liquid no reorientation occurred. The experiments of Taylor and of Saffman were made with spheroids of axial ratio about 2:1, and these results are probably most comparable to the behavior of stones in ice. Most other experiments have been made on very long thin specimens such as hair, paper fibres or fine glass rods (Eirich, Margaretha and Bunzl, 1936; Binder, 1939; Trevelyan and Mason, 1951). All of these workers find that Jeffery's equations correctly predict the behavior of the particles for a short time, but with a long time reorientation occurs. For particles with a small axial ratio (less than 15:1) Binder found that the reorientation was towards the transverse direction, while for greater axial ratios he found that, despite Jeffery's hypothesis, the parallel direction was that to which all particles tended. Eirich, Margaretha and Bunzl also observed that the particles with smaller axial ratios were more inclined to take up the transverse orientation.

To sum up this discussion, objects placed at random in a flowing liquid soon develop a parallel orientation of their long axes, and then, as time proceeds, they slowly develop a transverse orientation if their axial ratios are less than 15:1. Ice is not a linear viscous liquid, though the way in which it differs is not necessarily that assumed by Saffman. For ice the relation between shear stress and shear strain rate is not linear (Glen, 1955), and it is not yet known whether ice possesses cross-viscosity to any marked extent. Jeffery's theory will not therefore apply quantitatively to ice, but it is to be expected that the qualitative deductions will still be quite accurate, as they are based on rather general properties of a flowing material.

Collisions between stones.—Interactions between stones may influence the distribution of long axes. It is not simple to calculate what effect such interactions will have, as the effect of two stones interacting with each other will depend rather critically on the shape of the stones and the exact nature of the collision. Collisions may aid the stones to move from orbit to orbit towards the orbit of minimum energy, and in this way could favor the transverse orientation. In support of this suggestion is the observation of Manley, Arlov and Mason (1955) that the orbit of a single rod remained constant for a long time, whereas when the density of particles was increased the transverse orientation developed. This has led Manley, Arlov and Mason (1955, see also Manley and Mason, paper to be published) to suggest that the reorientation is primarily a collision effect.

A reorientation due to collision is to be expected on grounds other than those connected with Jeffery's minimum energy hypothesis. A stone which is rotating in a given orbit has a chance of colliding with another stone which is larger the larger the volume swept out by the stone in rotating in the orbit. Stones have, therefore, less chance of colliding, and so stay longer in an orbit, if the volume swept out in that orbit is small. Thus, if the main effect of collisions is to move stones from orbit to orbit at random, the effect of collisions will be to favor those orbits which sweep out the least volume. This argument is rather similar to that employed to demonstrate the development of the parallel peak during free shear; the stone spends the longest time in the orbit which sweeps out the least volume. For a prolate stone the orbit which sweeps out the least volume is that in which the long axis is transverse, so that the stone rotates about it. Thus for prolate stones collisions favor the transverse orientation for the long axis, and this happens to be the orientation favored also by the minimum energy hypothesis. For oblate stones, however, the orbit sweeping out least volume is that in which the *smallest* axis is transverse so that the stone rotates about that. This orbit is that of maximum energy dissipation for oblate spheroids, so that for oblate stones the minimum energy hypothesis still favors the transverse peak, while the collision probability favors the parallel peak.

For a stone which is neither prolate nor oblate (i.e. whose intermediate axis is not equal either to the shortest or the longest axis) the orientation favored by collisions will be transverse if the intermediate axis is more nearly equal to the shorter axis (prolate type stone) and will be parallel if the intermediate axis is more nearly equal to the longest axis (oblate type stone). The former case follows directly from the argument above, while for the latter case, the orbit sweeping minimum volume will be that in which the shortest axis is transverse, i.e. in which both longest and intermediate axes are in the plane of shear, and for this orbit the rate of rotation will vary, being least when the long axis is in the parallel direction, which is thus the most probable orientation.

The argument we have used here suggests that due to collisions the stones showing the transverse peak should statistically be more prolate in form than those in the parallel peak. It is perhaps significant that Holmes (1941) found "a slight, though gradual, increase in preference for parallel orientation with

decreasing difference in axial lengths". Holmes is here referring to the difference in lengths between the longest and the intermediate axes so that a small difference in length corresponds to an oblate type stone.

Interaction with the bed or across a shear plane.—The next type of interaction that we must consider is that between particles embedded in the moving ice and the stationary layer beneath. This layer may be bed rock, but it is more likely to be deposited till or perhaps even another ice layer, if the moving ice is overriding dead or slower moving ice along a shear plane. As Holmes has pointed out, a stone which projects to the bottom of the moving layer will be dragged along the layer underneath. This will tend to orient the stones in the parallel direction and may also reorient in this direction stones in the underlying layer. Holmes discussed this process in some detail and considered it the primary mechanism by which the parallel orientation is produced; for that reason we will not enter into great detail here. It is obviously a powerful method of orientation, and will, as Holmes has pointed out, tend to orient the long edges of stones rather than their long axes; thus for wedge-shaped stones one of the two sides of the wedge will tend to be parallel to the flow rather than the bisector of the angle. This is confirmed by the observation that wedge-shaped stones tend to have striae parallel to the sides of the wedge.

In confirmation of this type of orientation we can note the observation of Eirich, Margaretha and Bunzl that when their suspension of cylinders in a liquid was forced down a tube, the cylinders near the wall soon assumed a parallel orientation, whereas near the axis of the tube (where there is no strong shearing) the cylinders remained in apparently random orientations.

Deposition of till from the ice.—The process of deposition of the till may well affect the preferred orientation developed during flow. If till is deposited from active ice, so that material is added onto the top of a growing layer, the principal effects will be those due to interaction between the moving ice and the stationary till layer. As we have just seen, this tends to emphasize the parallel orientation. Holmes has pointed out another effect that may enter here. If a stone is rotating about an axis which is not one of circular symmetry, then the theory of Jeffery shows that the longest direction in the plane of shear spends comparatively little time near the perpendicular direction, and so has a preferred orientation which is horizontal. However, the chance of a collision with a stone above is highest when the axis is near to perpendicular, for it is then projecting into a region of much faster moving ice. If while it is rising into the perpendicular direction, a stone above, and thus moving faster than our stone, interacts with it from above, it will tend to force our stone downwards, and thus may push it into the till layer and bring it to rest. Thus in active ice the perpendicular orientation is more favored than would be predicted from the flow orbits alone. When the long axis is transverse, the same argument can be used to predict a greater number of perpendicular intermediate axes than are expected during flow, and Holmes explained in this way his observation that the class of stones with steeply inclined intermediate axes had a larger percentage of long axes in the transverse direction than had the class with low intermediate axis dip.

When till is formed from dead ice the effects will be rather different. Apart from bodily movements of the dead ice mass, the principal effect will be that, as the ice melts, the stones will fall down onto the material below. This will tend to drop the stones down so that their long axes are parallel to the bed (usually approximately horizontal), but will not affect the orientation distribution of long axes projected onto the parallel plane. It is for this reason that the rose diagram is probably a more satisfactory indicator of processes connected with ice flow than the contoured equal area diagram. In the rose diagram the number of stones whose long axes project onto the horizontal within a given angular range is plotted on polar co-ordinates against the angle, and this will be affected by deposition only to the extent that stones twist about a vertical axis as they are released from the melting ice. If the dip of the till stratum is large, then it is preferable to plot rose diagrams for the plane of the stratum rather than for the horizontal, but unless the angle is greater than about 30° no great errors will enter if this refinement is ignored. Contour diagrams, in which the dips as well as the horizontal angles of the long axes are recorded, will be much influenced by the way in which a stone drops onto the material beneath it.

Disturbance of till after deposition.—After till has been deposited from the ice further influences may upset its preferred orientations. Slumping of the whole deposit due to the melting of dead ice beneath may cause complications in the pattern, especially if the deposit itself flows during such slumping. For example a deposit on the edge of a kettle hole may flow into the hole and so obtain a preferred orientation in the direction of flow into the hole. Another process of this sort has been suggested by Hoppe (1952) to account for the observation that in hummocky moraine regions the preferred orientation of the stones is frequently perpendicular to the direction in which the moraine ridges run, even when these form curved loops. He suggested that till is pushed by the weight of overlying ice up into subglacial crevasses and in this way acquires its preferred orientation.

Solifluxion after the retreat of all the ice from the neighbourhood of the till may also be expected to change the preferred orientation, particularly if it is continued for any time, and if a mass movement of the deposit results. None of these effects will in general be related to the old direction of ice flow, and in any case in which stone orientation measurements are used to determine directions of ice flow, care must be taken to select deposits that do not appear to have been seriously disturbed since deposition.

SUMMARY OF EFFECTS PRODUCING PREFERRED ORIENTATIONS

We can now summarize the processes which have been discussed in the last section.

Preferred orientation parallel to the direction of flow: is produced by

free flow in the ice,

stone collisions (oblate stones only),

dragging on a stationary layer or over a plane of discontinuous shear (particularly during deposition from active ice).

Preferred orientation in the transverse direction: is produced by
protracted free flow in the ice,
stone collisions (prolate stones only),

Preferred orientation in the perpendicular direction: is produced by
stones being forced into a stationary layer by faster moving material above.

Preferred orientations unrelated to the direction of ice flow: is produced by
slumping of stones during deposition from dead ice (tending to eliminate large angles of dip),
subsequent flow of the till due to the melting of dead ice or extrusion from under masses of dead ice,
solifluxion.

COMPARISON OF PREDICTIONS WITH OBSERVED STONE ORIENTATIONS AND DEDUCTIONS WHICH CAN BE DRAWN FROM ORIENTATION MEASUREMENTS

We have already mentioned several ways in which the theory of stone orientations in till deposits is confirmed in practice. The majority of undisturbed deposits show either one or two directions in which the orientations have a maximum on a rose diagram. In most the largest peak is parallel to the direction of ice flow. Where there is a second peak it is usually at right angles to the first, i.e. transverse to the direction of ice flow. The second peak may exceed the first in magnitude. In no case does the transverse peak exist by itself. These observations alone show how well the stone orientations agree with predictions. There are probably significant correlations between prolateness and the tendency to have a transverse peak, which we would account for by the effect of collisions between stones. What remains to be explained, however, is why some rose diagrams lack the transverse peak. Two possibilities can be considered. Either the stones have not been in the ice long enough to develop the transverse peak, or they may have had it and lost it during the deposition process. In the former case the stones must have travelled only a short distance actually entrained in the ice, though they may have travelled further while being dragged beneath the glacier proper, while in the second case the deposition process must have involved a protracted period of dragging of the material.

Some confirmation for this approach is obtained from observations on tills in process of formation at Brageneset, Nordaustlandet (Donner and West, 1956). Here it was possible to measure stone orientations in tills which, although free of ice themselves, had ice both above and below them. Large transverse peaks were found only in narrow band tills (cf. diagram 3 in fig. 4), whereas in the thicker deposits only a single major peak was observed (diagrams 1 and 2 in fig. 4 show the extreme cases obtained). If the explanations suggested above are accepted, this would imply that the thick band tills were formed by a plastering-on process of lodgement which effectively eliminated the transverse peak, or that the stones in this till had never been completely immersed in flowing ice for a long time, while the narrow band tills were probably freer from the sliding process in formation, and, being relatively freer to flow with the ice, developed the transverse peak by free flow or as a

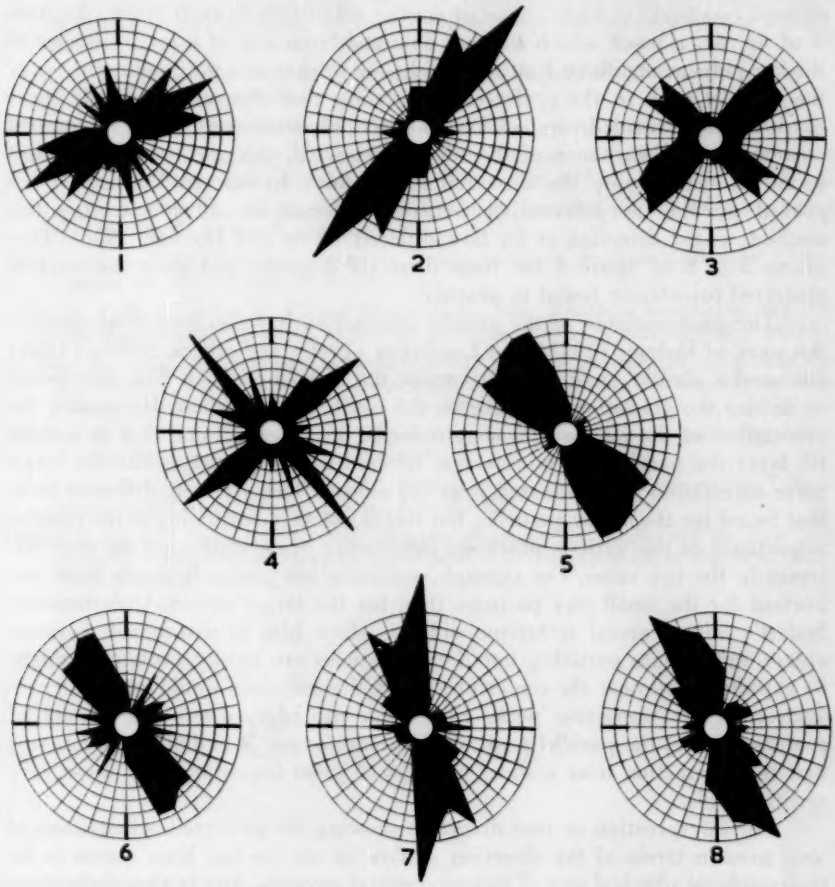

Fig. 4. Rose diagrams showing long-axis orientation of stones in various tills. The four cardinal points of the compass are shown on the diagrams, magnetic north at the top. The diagrams are from the following tills:

1 and 2. thick lodgement tills, Brageneset, Nordaustlandet, Spitsbergen.

3. narrow band till in ice, Brageneset, Nordaustlandet, Spitsbergen.

4. narrow band till from the Contorted Drift on the Cromer coast near Sheringham, Norfolk, England.

5 to 8. various lodgement tills in eastern England.

result of collisions during free flow. It would be dangerous to generalize from this one place and to state that transverse peaks are always associated with narrow band tills, but this experience does show that in the case of the Brageneset deposits there was a qualitative difference in morphology between the deposits with large transverse peaks and those without. The principle may have validity in other locations, however, as observations of stone orientations from a narrow band till in the Cromer deposits (about 20 cm wide) also

showed two peaks at right angles of similar magnitude to each other (diagram 4 of fig. 4), a result which was not obtained from any of a large number of thicker lodgement tills in East Anglia. In the thicker tills the predominance of the parallel peak in the preferred orientation rose diagram is usually quite marked, and, provided care is taken to select only sites in which no subsequent movement or disturbance of the till has occurred, this method can be most useful in determining the direction of ice flow. In eastern England it has proved possible to differentiate between different ice advances using this method, as the direction of ice flow differed (West and Donner, 1956). Diagrams 5 to 8 of figure 4 are from these till deposits, and show the kind of preferred orientation found in practice.

The predominance of the parallel orientation has also been confirmed by the work of Holmes (1941) and Lundqvist (1948) and others. Seifert (1954) has used a similar method to determine the direction of ice flow, and hence to deduce the glaciation sequence on the north German coast. He studied the orientation of the microscopical particles in the till. He found that in a thick till layer the parallel orientation was favored near the edge, while the transverse orientation predominated near the centre. This result is different from that found for the larger particles, but this is not very surprising as the relative importance of the various processes influencing orientation must be very different in the two cases. For example, collisions are probably much more important for the small clay particles than for the larger stones. Unfortunately Seifert's microscopical technique did not allow him to distinguish between oblate and prolate particles, but if the particles are mostly prolate, it might be expected that, near the center of a band, the collisions would assist the production of the transverse peak, while near the edge, where the density of particles is less, the parallel peak would predominate. It is difficult to imagine that the interaction over a shear plane is of great importance for these very small particles.

The interpreation of rose diagrams showing the preferred orientations of long axes in terms of the direction of flow of the ice has been shown to be reasonable on physical as well as experimental grounds. Any further deductions made from the nature of the diagrams such as the relative magnitude of the transverse peak, are somewhat more dubious, owing to the large number of processes that could have acted to give the peaks their relative magnitudes. It would seem, therefore, that stone orientation measurements are most valuable for determining the direction of ice flow, but less significant for other purposes.

ACKNOWLEDGMENT

We would like to thank Mr. P. G. Saffman for permission to quote from his unpublished work, and for his help in finding the literature references on the behavior of suspensions of objects in liquids.

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SEDIMENTARY CHARACTERISTICS OF DUST STORMS: III. COMPOSITION OF SUSPENDED DUST

W. S. CHEPIL

ABSTRACT. Wind-blown dust varied widely in its composition depending on the composition of eroded soil, the year of measurement, and the distance and height of transport. The composition of the dust was like the composition of many samples of loess. The size distribution of dust and of coarser materials transported at any height or deposited anywhere after any single windstorm was characterized by a single peak diameter of the discrete particles and by arms on each side of the peak falling off independently of each other at some constant rate. The peak diameter varied from one graded material to another, depending on the physical nature of the soil, distance and height of transport, and possibly the velocity of the wind.

INTRODUCTION AND ACKNOWLEDGMENT

In an earlier study the grading patterns of soil materials transported by wind and ultimately deposited in drifts here and there in the vicinity of eroded fields were presented (Chepil, 1957). Considerable proportions of fine soil particles known to be transported by wind were deposited nowhere in the vicinity of wind-eroded localities. These particles were suspended in the form of dust clouds and transported far and wide by the atmosphere. A study was made on the physical characteristics of the suspended dust and of some of the materials transitional between dust and drift. The results of this study are presented herein.

The object of this study was (1) to obtain an insight of the nature of sorting of fine particles from the less mobile portions of the soil and (2) to determine whether dust carried in the atmosphere is comparable in composition with loessal soils.

This paper is Contribution No. 549, Department of Agronomy, Kansas Agricultural Experiment Station, Manhattan, Kansas, and Agricultural Research Service, U.S.D.A., Cooperative Investigations on the Mechanics of Wind Erosion.

PROCEDURE

Dust was caught at various heights above the ground in accordance with a method already described in an earlier publication of this series (Chepil and Woodruff, 1957). The equivalent size distribution of the dust particles was determined in accordance with a method also described in the earlier publication. Mechanical composition of the dust, on the other hand, was determined by the method of Bouyoucos (1951).

SIZE DISTRIBUTION AND MECHANICAL COMPOSITION OF DUST

The equivalent size distribution of dust particles carried in suspension is shown in the "percentage smaller" diagrams of figure 1. Since the size distribution curves appear to be quite symmetrical, the mean equivalent diameter of the particles can be determined approximately from a point on the distribution curve corresponding to a percentage smaller value of 50. Figure 1 indicates that for both 1954 and 1955 the mean equivalent diameter of discrete dust particles (shown on curve a) was considerably greater than the average diameter of elementary mechanical particles (sand, silt, and clay,

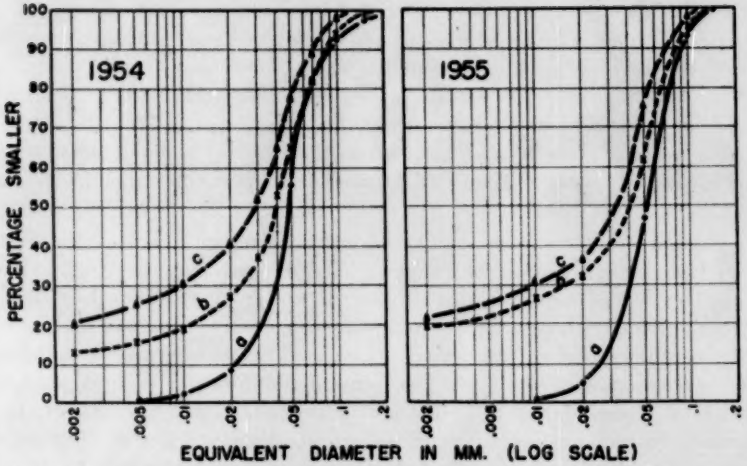

Fig. 1. Average equivalent size distribution of soil particles carried in suspension between 4 and 8 feet height during 1954 dust storms and between 2 and 20 feet height during 1955: (a) discrete particle size distribution as determined by sedimentation in carbon tetrachloride, based on a composite of all samples, (b) mechanical composition of dust after dispersion with sodium hexametaphosphate in water, composite of samples over sandy soils, and (c) composition as in (b) for a composite of samples over silt loam soils.

shown on curves b and c) of which the dust particles were composed. For the 1954 data, the mean equivalent diameter of the dust particles over all of the soil classes investigated was 0.049 mm. The mean equivalent diameter of the mechanical particles, on the other hand, was 0.039 mm for samples collected over sandy soils and only 0.029 mm for samples over silt loam soils. The values of mean equivalent diameter for 1954 were only slightly smaller than the corresponding values for 1955.

It is evident that some dust particles were composed of aggregates of sand, silt, and/or clay, otherwise there would not have been any dispersion of the dust particles by the dispersing agent in water and the size distribution curve of the elementary mechanical particles would have coincided with the size distribution curve of the discrete dust particles. The greater the spread between the two average curves, the greater would be the degree of aggregation of the elementary mechanical particles. It is shown from figure 1 that the greatest spread, and therefore the greatest degree of aggregation, was for the clay particles (< 0.002 mm), less for silt ($0.002 - 0.05$ mm), and least for sand (> 0.05 mm). It is evident that the largest dust particles were almost entirely composed of individual particles of fine and very fine sand. Considerable deviations between the two curves within the silt and clay range seem to indicate that silt and clay were carried to some degree as aggregates of the size of silt and sand. Furthermore, curve a indicates that virtually all of the clay fraction was aggregated into particles greater than 0.05 mm in diameter. Photographs of dust particles substantiate this conclusion. Although discrete particles of the size of clay amounted to a mere trace of the total weight of

the material, their numbers were fairly substantial, especially in some of the finest dust samples. The great majority of the dust particles were larger than particles of clay, were smooth, rounded, and almost transparent, and composed predominantly of individual grains of silt and sand. An occasional dust particle showed some smaller particles clinging to its surface. Still others showed definite angularity and unevenness of surface. The latter particles apparently were aggregates composed primarily of clay, very fine silt, and organic matter.

Although the percentage smaller diagrams of figure 1 serve a good purpose of indicating the general range of size of dust particles, they do not indicate clearly the true size distribution of the particles. Therefore, it was decided to re-plot the data by following the log diagram method of Bagnold (1943) originally used on dune sand. This was done by plotting the logarithm of the percentage weight of soil per unit of log-diameter scale against the logarithm of grain diameter. Assuming the diameter limits of each grade as determined by sieving to be d_1 and d_2, the logarithmic interval of grain diameter is given by log d_1/d_2 which for convenience is designated by D. The percentage weight p of each grade per unit of log-diameter scale is then equal to $\Delta p/\Delta D$. These percentage frequency values plotted against the logarithm of particle diameter indicated a single "peak" or predominant diameter of the dust particles. Moreover the grades to the left and the right of the peak diameter fell off at rather uniform rates (fig. 2, curve a). Also, the peak diameter coincided with the mean equivalent diameter of the dust particles (table 1). The size distribution pattern of the dust particles was of the same general nature as the size distribution of dune sand (Bagnold, 1943) and of

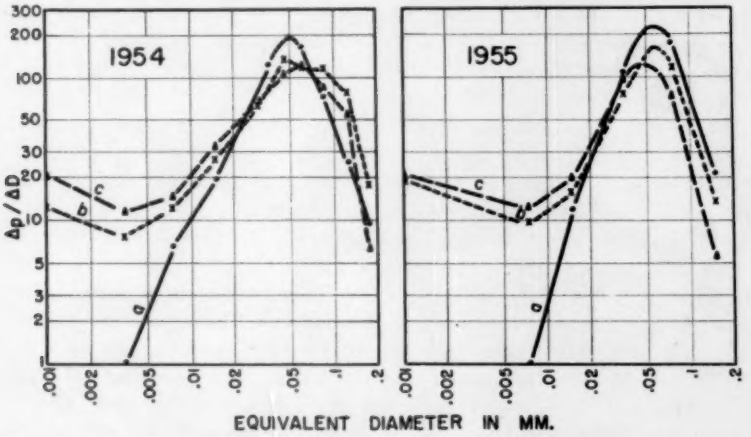

Fig. 2. Average equivalent size distribution of soil particles carried in suspension between 4 and 8 feet height during 1954 dust storms and between 2 and 20 feet height during 1955: (a) discrete particle size distribution as determined by sedimentation in carbon tetrachloride, based on a composite of all samples, (b) mechanical composition of dust after dispersion with sodium hexametaphosphate in water, composite of samples over sandy soils, and (c) mechanical composition as in (b) for a composite of samples over silt loam soils.

drifted soil materials (Chepil, 1957). The only distinguishing features of all the distribution curves were in their positions on the logarithmic scale and in the slopes of each of the two straight arms. It is evident that the same logarithmic law of distribution applicable to sand and soil grains transported in saltation and surface creep is also applicable to suspended dust.

TABLE 1

Equivalent Diameter Distribution and Mean Equivalent Diameter of Dust Particles Carried at Different Heights Over Two Soil Class Groups During 1955 Kansas and Colorado Dust Storms.

Class	Height	Equivalent diameter distribution					Weighted mean equivalent diameter
		0.25-0.1 mm	0.1-0.05 mm	0.05-0.02 mm	0.02-0.01 mm	<0.01 mm	
	feet	%	%	%	%	%	mm
Sandy	2	15.0	64.5	19.6	0.9	0	0.082
soils	5	2.7	43.6	46.1	7.6	0	0.054
	11	3.3	42.1	50.4	4.3	0	0.055
	20	0.8	39.2	56.5	3.5	T	0.051
Mean		5.4	47.4	43.1	4.1	T	0.060
Silt	2	21.3	63.9	14.2	0.6	0	0.090
loam	5	7.8	49.1	39.4	3.7	0	0.065
soils	11	2.0	37.0	51.8	9.2	T	0.050
	20	1.6	31.1	63.6	3.7	T	0.050
Mean		8.2	45.3	42.2	4.3	T	0.064

The average equivalent size of dust particles carried in suspension between the heights of 2 and 20 feet was considerably smaller than the average equivalent size of particles of accumulated drifts analyzed in the previous study (Chepil, 1957, tables 1, 2, and 3). Analysis of equivalent size distribution of particles in the drifts and of particles carried through the air revealed that considerable proportions of intermediate sizes were absent from the two materials. The intermediate sizes were evidently carried predominantly below the height of 2 feet and transported beyond the vicinity of the drifts. They were apparently transitional materials carried partly in saltation and/or partly in true suspension. Comparative differences between soil particles accumulated in the drifts, the transitional particles carried beyond the vicinity of the drifts, and the dust particles carried between the heights of 2 and 20 feet are shown in table 2.

The size distribution of particles collected from the drifts and from the air was characterized by a single peak diameter with arms to the left and the right of the peak falling off each at its own constant rate. The transitional material bore no similarity to such a grading pattern. Nowhere could be found deposited or suspended materials bearing any similarity to this missing material. Evidently the missing material was deposited nowhere in particular but rather constituted parts of materials ultimately deposited at various locations from their source. An example of one such part is indicated as a lee deposit in table 3. The lee deposit was composed of particles generally much finer than those found in drifts or mounds in or near an eroded field; but on the

other hand, it contained particles substantially coarser than those generally caught in the air between the heights of 2 and 20 feet.

TABLE 2

Equivalent Size Distribution of Soil Particles in Fields, in Accumulated Drifts, in Areas Beyond the Vicinity of the Drifts, and in the Air Between the Heights of 2 and 20 Feet.

Material	Equivalent diameter of particles						
	>1.38 mm	1.38-.58 mm	.58-.30 mm	.30-.15 mm	.15-.073 mm	.073-.037 mm	<.037 mm
	%	%	%	%	%	%	%
Original soil	21.5	8.4	21.9	25.2	11.0	7.2	4.8
Drifts or dunes	0.5	7.2	35.8	34.2	14.0	5.7	2.6
Eroded but not found in drifts (estimated from composition of original soil)	0	0	T	29.4	25.4	29.1	16.1
Caught in air between 2 and 20 feet	0	0	0	1.1	5.6	40.2	53.1

TABLE 3

Size Distribution of Soil Particles in Drifts and in a Lee Deposit Originating from a Wind-eroded Field of Loess Soil near Salina, Kansas, 1950.

Source of material	Diameter of particles					
	2.0-0.84 mm	0.84-0.42 mm	0.42-0.21 mm	0.21-0.1 mm	0.1-0.05 mm	<0.05 mm
	%	%	%	%	%	%
Drifts in and nearby eroded field	2.3	30.2	38.6	16.7	2.9	9.3
Uniformly diminishing deposit in Sudangrass beyond 400 feet to the lee of the eroded field	0	1.0	8.2	34.3	22.4	34.1

The size distribution of particles of the lee deposit, as in all graded soil materials, was characterized by a peak diameter with each of the two arms falling off uniformly but independently of the other. The lee deposit varied from nearly zero to over 2 inches in depth depending on distance from the eroded field. The deposit had no tendency to form drifts or dunes like those formed from coarser materials, but was distributed rather uniformly over the ground well protected by a cover of Sudangrass. The absence of a grass cover no doubt would have caused the material to be moved rapidly and distributed somewhere else by the wind.

The size distribution of the elementary mechanical fractions showed a distinctly different pattern from the size distribution of the dust particles com-

posed of those fractions. The size distribution curves indicated a major bulge in the vicinity of the peak diameter of the dust particles and a small rise in the clay fraction (fig. 2, curves b and c). The predominant equivalent diameter of the mechanical fractions of which the dust particles were composed varied from about 0.05 to 0.07 mm depending on soil class and year of measurement. The corresponding mean equivalent diameter of the mechanical fractions, on the other hand, was only 0.048 mm for sandy soils and 0.035 mm for silt loam soils (table 4). These discrepancies are due no doubt to the fact that the mechanical fractions were not symmetrically graded. The dust particles, on the other hand, were symmetrically graded so their peak diameter (fig. 2, curve a) coincided with their mean equivalent diameter (table 1).

The mean equivalent diameter of the suspended dust particles decreased with height above the ground (table 1). Particles carried at 2 feet above the ground had an average equivalent diameter of 0.086 mm whereas those carried at 20 feet had an average equivalent diameter of 0.05 mm. As determined by extrapolation in the previous study for 1955, the average equivalent diameter of dust particles at one-mile height was only 0.022 mm (Chepil and Woodruff, 1957). The equivalent size of particles carried over sandy soils and over silt loam soils at corresponding heights was about the same. This similarity of equivalent size is to be expected since the same wind force acted on all of the soil classes. The actual size of particles was conceivably different for the various soil classes, depending on their bulk density.

Dust arising from sandy soils contained more sand and less silt and clay than dust carried at corresponding heights over silt loam soils (table 4). However, samples of dust from sandy soils were similar in composition to dust carried at some lower height from silt loam soils. For example, dust carried at 11 feet over sandy soils differed little in mechanical composition from dust carried at 2 feet over silt loam soils. It is reasonable to expect that dust carried at greater heights would be carried to greater distances. If this assumption is true, then some deposits of dust originating from sandy soils would be similar in composition to some deposits derived from silt loam soils, but located farther away from their source.

DISCUSSION AND CONCLUSIONS

The wind-blown dust obtained in this study (fig. 1, curves b and c, and table 4) generally contained somewhat more sand and clay than the samples of loess analyzed by Péwé (1951) and the samples of wind-blown dust analyzed by Swineford and Frye (1945). On the other hand, the clay content of the dust obtained in this study was generally much lower than the clay content of wind-blown dust collected by Warn and Cox (1951). This study shows that considerable variation exists in the composition of suspended dust depending on the composition of eroded soil, year of measurement, and distance and height of transport. Differences obtained in different years probably are associated with different wind velocities. Considerable variations do exist in the composition of loess. For example, Hanna and Bidwell (1955) showed that the elementary particles of loess become finer with distance away from the Missouri River valley flood plain believed to be the source of loess. The com-

TABLE 4

Mechanical Composition of Dust Carried in Suspension at Different Heights over Two Soil Class Groups During 1955 Kansas and Colorado Dust Storms.

Class	Height	Fine sand 0.25-0.1 mm	Very fine sand 0.1-0.05 mm	Coarse silt 0.05-0.02 mm	Medium silt 0.02-0.01 mm	Fine silt 0.01-0.002 mm	Clay <0.002 mm	Weighted mean equivalent diameter mm
	feet	%	%	%	%	%	%	
Sandy soils	2	10.5	36.6	24.8	5.0	7.6	15.5	0.068
	5	2.0	37.3	29.7	5.0	7.1	18.9	0.043
	11	3.3	34.7	27.7	5.6	7.1	21.6	0.043
	20	0.7	30.6	37.5	3.0	5.6	22.6	0.038
Mean		4.1	34.8	29.9	4.6	6.9	19.7	0.048
Silt loam soils	2	3.0	31.6	29.3	6.1	9.8	20.2	0.041
	5	1.4	19.5	43.7	6.1	8.4	20.9	0.034
	11	1.5	20.7	41.8	6.0	8.4	21.6	0.034
	20	0.8	19.9	41.4	5.9	8.4	23.6	0.032
Mean		1.7	22.8	39.1	6.0	8.8	21.6	0.035

position of dust obtained in this study is like the composition of many samples of loess.

Results of analysis of suspended dust and of some deposited materials agree in many respects with previous results on lag materials and drifts or dunes and confirm the existence of a remarkable phenomenon of sorting of soil materials by the wind, namely: soil material transported at any height or deposited anywhere by any single windstorm is characterized by a predominant or peak diameter of the discrete particles and by arms on each side of the peak falling off each at its own uniform rate. The peak diameter varies from one graded material to another depending on the physical composition of the soil and distance and height of transport. Materials deposited at a location by different winds are composed of mixtures of different grades, may have more than one distinct or several indistinct diametric peaks, and show no resemblance to any single grade. The finer the eroded particles the greater is their speed, height, and distance of travel. The individual grades, such as the nonerodible fractions, the lag materials, the drifts or dunes, the lee deposits, and the dusts tend to merge one with the other so that no clear-cut demarkation is found between any of them. The grades are merely descriptive and not strictly categorical.

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ANISOTROPY OF FRACTURE IN QUARTZ

F. DONALD BLOSS

ABSTRACT. Specimens of "rock crystal," rose, and milky quartz were crushed both at room temperatures and in the 650-700°C range. Slide mounts of the 100-200 mesh fragments were examined on the universal stage to determine for each fragment θ, the angle between the fragment's optic axis and the normal to its plane of rest. Histograms, based on several thousand such measurements, illustrate the relative frequencies of observed θ values.

Study of these histograms indicates a definite crystallographic control of fracture. Maxima near or at 52° are interpreted as the expressions of cleavage parallel to the *r* and *z* rhombohedra of low quartz and to the corresponding unit dipyramid of high quartz. The greater height of this peak for the specimens crushed at 650-700° (as compared with those crushed at room temperature) may be interpreted either as the result of the high quartz possessing better cleavage than low quartz for this direction or as the result of ruptures developing parallel to this direction when the specimen was heated through the inversion temperature. Lesser maxima at 90°, present for both temperature modifications, indicate prismatic cleavage is next in importance. Comparison of these peaks at 90° for high and low quartz indicates that the prismatic cleavage of low quartz is superior to that of high quartz.

The histogram representing the frequency distribution of θ that would be secured if there were complete isotropy of fracture orientations is actually sinusoidal. It does not divide the data equally among all classes of θ; instead the classes possess increasingly greater frequencies as the values of the class limits of θ become larger.

The histograms for crushed quartz generally indicate a small peak in excess of the isotropic distribution at 0°. For both temperature modifications this is interpreted as the result of the basal cleavage. The histograms possess some characteristics that indicate the existence of cleavage parallel to *x* and *s*, though the evidence is not conclusive.

Fractures, planar or otherwise, are relatively rare along planes making angles of 2° to 36° with the basal pinacoid. With few exceptions the isotropic distribution gives values greater than observed results within this region.

Deformation lamellae in quartz, which are most frequently oriented subparallel (within 2° to 36°) to the basal pinacoid, are not likely to represent healed fractures as is believed by some. In fact,the histogram compiled by Ingerson and Tuttle (1945) to illustrate the frequency with which various orientations of deformation lamellae occur bears a rather antipathetic relationship to a similar histogram for quartz fractures. This antipathetic relationship, plus the marked divergency between the Ingerson and Tuttle histogram and the theoretical isotropic distribution, suggests that there is a rather definite crystallographic control over the orientation of deformation lamellae.

INTRODUCTION

Several conflicting statements persist with reference to the cleavage tendencies of quartz. Winchell (1951, p. 246-7) states that high quartz has distinct $(10\bar{1}1)$ and $(10\bar{1}0)$ cleavages whereas in low quartz the cleavage is " . . . absent or difficult and poor parallel to $10\bar{1}1$, etc." Sosman (1927, p. 488) summarizes data for low quartz which indicate that cleavage occurs more readily parallel to the positive rhombohedron *r* than to the negative rhombohedron *z*.[1] Prismatic cleavage has been reported by Drugman (1939, p.

[1] The nomenclature for quartz crystallographic planes followed in this paper is:

c — the basal pinacoid, (0001).
m — the unit prism, $(10\bar{1}0)$.
a — the second-order prism, $(11\bar{2}0)$.
r — the positive rhombohedron, $(10\bar{1}1)$.
z — the negative rhombohedron, $(01\bar{1}1)$.
s — the second order trigonal dipyramid, $(11\bar{2}1)$.
x — the trigonal trapezohedron, $(51\bar{6}1)$.

259) and by Berman (Fairbairn, 1939, p. 363). Ingerson and Ramisch (1942, p. 596), however, doubted the existence of a prismatic fracture; if it existed, they believed it to be completely overshadowed by the more prominent rhombohedral cleavage. Sosman (1927, p. 488) cited observers who " . . . question whether low-quartz has any cleavage whatever. The occasional plane fractures may be only the result of gliding planes."

Mallard (1890) observed the r (10$\bar{1}$1) and z ($\bar{1}$011) cleavages to be developed most readily when pressure was applied normal to a thin plate of quartz cut parallel to a (11$\bar{2}$0) plane; of the two, r was judged the better by virtue of the superior light reflectance of the surfaces thereby formed. Fracture directions parallel to m (10$\bar{1}$0) and c (0001) were interpreted by Mallard to be distributive cleavage surfaces formed of combinations of r and z. Anderson (1945, p. 426) similarly concluded that r and z are common cleavage directions, but that ruptures parallel or subparallel to m, c, or a are also fairly abundant. The experiments of Griggs and Bell (1938) produced needles of quartz elongated parallel or subparallel to the c-axis, to the unit rhombohedra, and to the basal pinacoid. The relative abundance of these several needle-types was determined largely by the orientation of the direction of stress application with respect to the c-axis of the quartz cylinder examined. Recently, Borg and Maxwell (1956), in a study of 250 planar fractures developed in grains of a quartz sand subjected to a directed stress, concluded that the r and z rhombohedra represent the most important surfaces for rupture. Fracture planes parallel the prism zone, particularly m, were next in importance.

Fairbairn (1939) predicted from the structure of low quartz that the order of ease of cleavage development should be r, z, m, c, a, s, and x. The latter two were, he believed, improbable as independent cleavages.

The present study was undertaken in recognition of the divergency in the preceding opinions. The new experimental data clarify at least a few of the discrepancies. Since curved fractures are in effect included in the results, the study permits an initial assessment of the degree of anisotropy of fracture in quartz.

EXPERIMENTAL METHODS

Selection and description of specimens studied.—In order to determine whether type of quartz had influence on the fracture tendencies, three very different varieties of quartz were chosen for study. "Rock crystal" was represented by a large, clear, prismatic specimen, four inches in length, terminated at one end by unequally developed r and z faces. Massive specimens of milky quartz and rose quartz were used, the latter containing numerous elongate inclusions all approximately parallel to the c-axis.

Preparation of slide mounts studied.—Each variety of quartz was individually crushed in a jaw crusher and the 10-16 mesh sieve fraction of each was then split into two fractions. One was crushed by ramrod-like action while in the blocked-off segment of a steel pipe at room temperature; the other was similarly crushed after the pipe had been inserted into a tubular furnace and slowly heated until a temperature within the 650-700°C range had been main-

tained for 15-30 minutes. Temperature was determined by a pyrometer whose high-temperature junction was located near the material to be crushed. The ramrod-crushed material was sieved and the 100-200 mesh fragments were sprinkled upon glass slides and cemented in their rest positions with a hot, highly fluid, xylol solution of Canada balsam. Successively, richer Canada balsam solutions were applied and cooked to complete the mounting.

Preliminary examination of these slides showed that all grains rested at the bottom of the Canada balsam layer and presumably, therefore, were in contact with the slide at a minimum of three points. The check of the method using crushed calcite (to be described later) corroborated this.

In order to compare the effect of manner of crushing, a portion of the 10-16 mesh sieve fraction of crystal quartz was crushed in a hammermill rather than within the pipe. Aside from this variation slides were prepared precisely as before. This quartz, in contradistinction to the samples crushed in the pipe, is referred to as hammermill-crushed quartz.

Measurement of θ, the angle between the c-axis and the normal to the slide.—If at least three points of a grain's lower surface are in contact with the glass slide, θ (here defined as the angle between the *c*-axis and the slide normal) will represent the angle between the *c*-axis and the pole of the generalized fracture surface upon which the grain rests. For each grain, θ (or its complement) was measured with a 4-axis Leitz universal stage. Grains in contact were rejected to prevent measurements of fragments partly propped up by neighbors. Replications of measurements on quartz grains showed a precision of 2°; measurements are believed accurate to within 3°.

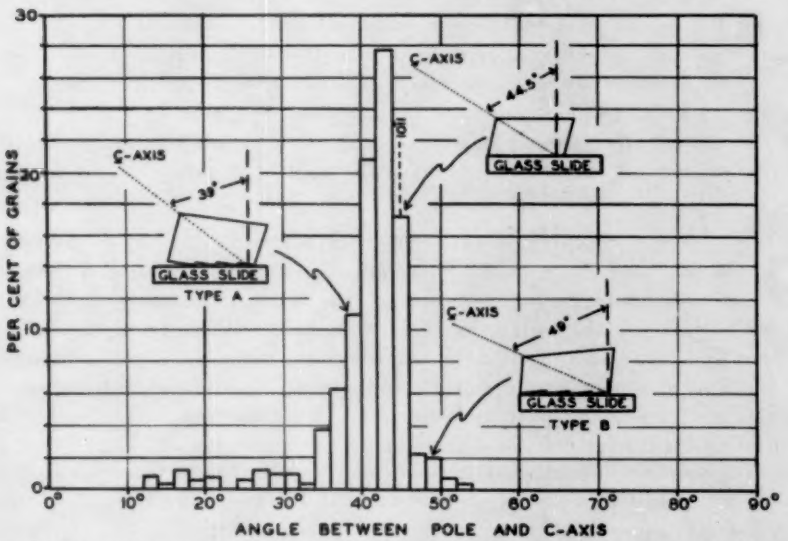

Fig. 1. Histogram showing relation between the *c*-axis of 200 grains of crushed calcite and the poles to the fracture surfaces upon which they rested. The two main types of distributive cleavage surfaces of calcite (labelled A and B) are compared with a true rhombohedral plane.

Test of method.—Iceland spar was hammermill-crushed at room temperature, slide mounts were made, and θ was measured for 200 untwinned fragments (using universal stage segments of refractive index 1.649). Expressed as a histogram (fig. 1), the results show an overwhelming concentration near 44.5°, the θ value one would expect to measure for a grain resting on a smooth (10$\bar{1}$1) plane. The validity of this method of examining a mineral for cleavage seems thus confirmed.

The unequal dispersion of observations around 44.5° may be explained by the relatively greater frequency of distributive cleavage surfaces of type A rather than of type B (fig. 1). Type B surfaces may be seen to possess the sharper re-entrant angles and therefore to be physically less likely to occur than distributive surfaces of type A. The disparity between the two types assumes even greater dimensions if one considers the nature of the histogram which represents isotropism (cf. the following section).

ISOTROPIC DISTRIBUTION OF FRACTURE SURFACES

Whereas the cleavage of calcite showed a very strong departure from isotropy in the test of the method here used (fig. 1), the cleavage tendencies of quartz, being so weak, invite comparison with a model for which complete isotropy of tendencies is assumed. Thus for a hypothetical uniaxial material possessing absolutely no preferred planar direction for fracture, the poles to the fracture surfaces would plot with equal density on a spherical projection (fig. 2). Such data *when presented on a histogram as the angle between fracture pole and c-axis* (as for the calcite data) *produce a distribution* (*fig.* 3)

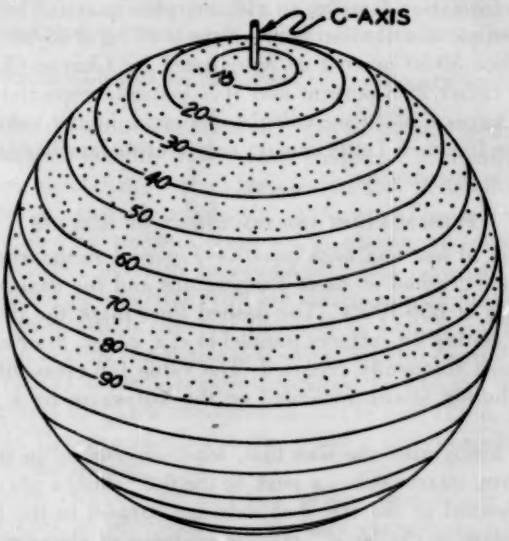

Fig. 2. Perspective sketch of the distribution of poles to fracture surfaces with respect to the *c*-axis for a uniaxial material possessing no cleavage. Note that the poles plot with equal density over the entire sphere.

which is clearly proportional to sine θ. P, the percentage of data of the isotropic distribution between the limits θ_1 and θ_2 may be calculated as:

$$P = 100 \int_{\theta_1}^{\theta_2} \sin \theta \, d\theta = 100 \, (\cos \theta_1 - \cos \theta_2). \qquad \text{Eq. 1.}$$

Thus the percentage of data (assuming a three dimensional isotropy of the fracture poles to the c-axis) can be calculated for any class of the histogram.

Equation 1 is not limited in application to fracture poles. It applies equally to *all* histograms which similarly express the angle between a given reference axis and a population of planes or lines which are distributed isotropically in three-dimensional space. The isotropic distribution of figure 3 is obviously not the horizontal line tacitly assumed in published histograms illustrating the preferred crystallographic orientation of elongations in sedimentary quartz

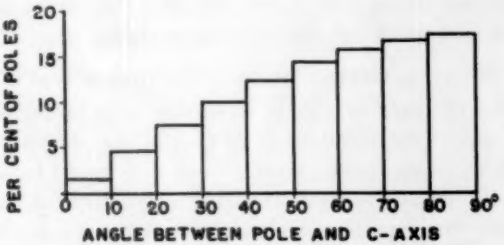

Fig. 3. Histogram showing relation between the *c*-axis and the poles to fracture surfaces for this hypothetical uniaxial material. Note the sinusoidal nature of the resultant isotropic distribution (10° class interval).

grains or of deformation lamellae in metamorphic quartz. The division of the data of the isotropic distribution between the 0-45° and 45-90° classes of such histograms is not 50-50 percent as anticipated by Chayes (Fairbairn, 1949, p. 301) but is rather 29.3 percent and 70.7 percent, respectively. Using these latter values, Chayes would have calculated a much higher value of chi square when testing the Ingerson-Tuttle data on deformation lamellae for compatibility with an isotropic distribution.

PRESENTATION AND DISCUSSION OF RESULTS

Comparison of low and high quartz.—Figure 4 summarizes observations for quartz grains crushed at room temperature and for grains crushed while at a temperature of 650-700°C. The dashed lines mark the theoretical values of θ for grains resting on surfaces parallel to *c*, *r*, *z*, *s*, *x*, *m* or *a*, the directions for cleavage most frequently reported. The value for every fifth class of the isotropic distribution is also indicated on the histogram by a short dash-dot-dash line.

From the histograms one sees that, whether crushed in the low or high temperature form, quartz shows a peak in the 0-2° class, a phenomenon probably best interpreted as the result of cleavage parallel to the basal pinacoid.

The maximum in the 56-58° class is evidence of cleavage parallel to the *r* and *z* directions in low quartz and parallel to the *r* direction in high quartz. The displacement of this peak to the right of 52°, the theoretical position, is

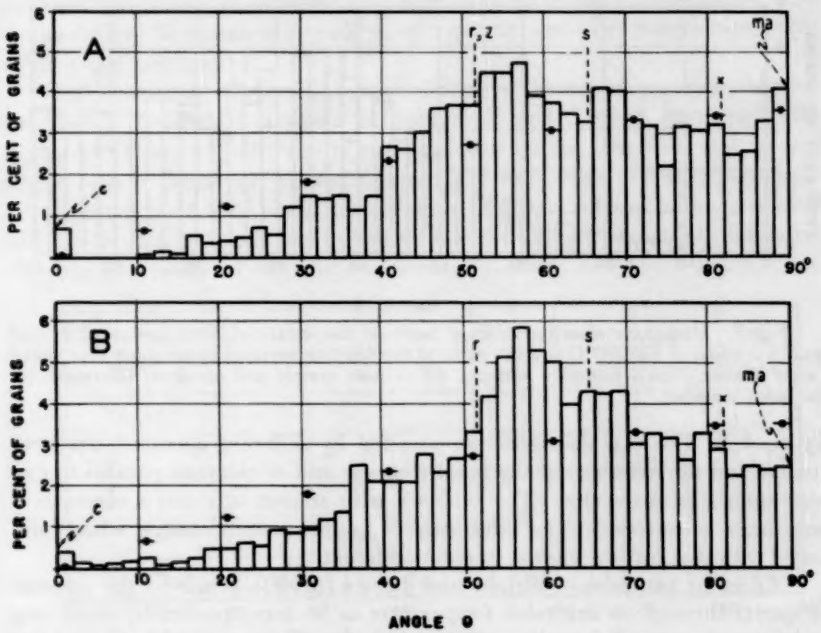

Fig. 4. A. Histogram showing relation between c-axis of 1500 grains of crystal quartz crushed at 25°C and poles to the fracture surfaces upon which they rested. The angles for *c*, *r*, *z*, *s*, *x*, *m*, and *a*, the most frequently reported cleavage forms, are indicated as dashed lines. The height of every fifth class of the isotropic distribution (2° class interval) is shown by a dash-dot-dash line in this and following histograms (i.e., figs. 5 to 10).

B. Similar histogram for 1500 grains of crystal quartz crushed at 650-700°C.

perhaps caused by the sinusoidal bias that the histogram method of plotting imparts to the data, or possibly by a greater frequency of distributive cleavage surfaces involving combination of rhombohedral and prismatic planes rather than combination of two rhombohedral surfaces. (The latter combination would be represented by surfaces possessed of sharper re-entrant angles and would probably be less likely to be produced.) The increased height of the peak for high quartz as compared with low quartz may indicate that: (1) the *r* direction in high quartz is superior to either or both *r* and *z* directions in low quartz; (2) in heating the quartz to 650-700°C prior to crushing, cleavages parallel to the $(10\bar{1}1)$ directions were developed by passage through the inversion temperature; or (3) a combination of (1) and (2) is involved.

Evidence for cleavage parallel to the *s* and *x* directions is not clear cut for either high or low quartz.

For low quartz cleavage parallel to a prism (dominantly *m* rather than *a*, according to work in progress) is strongly indicated. In high quartz this prismatic cleavage is less strongly developed, if at all.

Figure 5, representing results for high quartz, is reproduced by the kind permission of Mr. Louis Walter (1955), and permits a comparison with

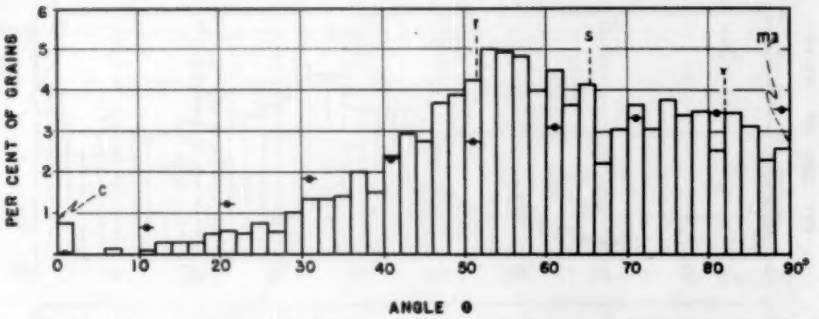

Fig. 5. Histogram showing relation between the c-axis of 1000 grains of crystal quartz crushed at 650-700°C and the poles to the fracture surfaces upon which they rested (after Walter, 1955). Compare with fig. 4B to note sample and operator differences for the same material.

figure 4 B. The two distributions compiled by different operators are very similar and the existence of the basal cleavage and of cleavage parallel to *r* in high quartz is again clear. The evidence with respect to *s* and *x* cleavage is once more inconclusive. The inferiority of the prismatic cleavage, when compared with that for low quartz, is again apparent.

Effect of inversion.—Wright and Larsen (1909) observed the passage of quartz through its inversion temperature to be accompanied by shattering and fracturing parallel to the unit rhombohedra. This occurred both on heating above and again upon cooling below the inversion temperature. Frondel (1945, p. 455) noted that both rate of heating or cooling and thickness of quartz affected the amount of such shattering and fracturing.

To test the influence of this effect on the results of figures 4B and 5, a standard 10-16 mesh charge of quartz grains was heated to 800°, then cooled to room temperature prior to crushing. Results shown in figure 6 indicate an

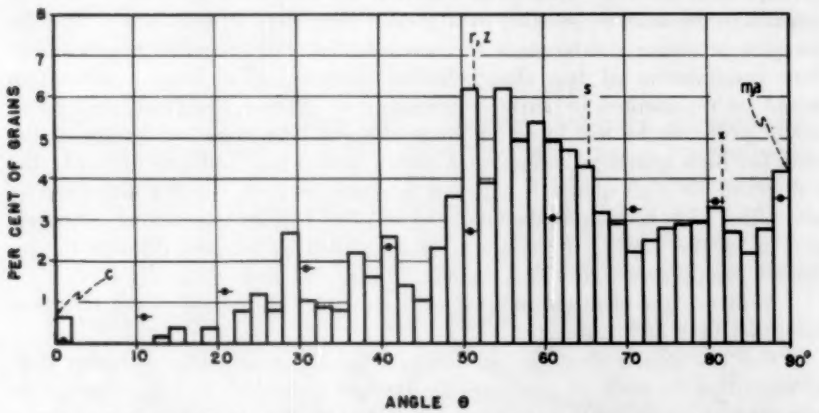

Fig. 6. Histogram showing relation between the c-axis of 500 grains of crystal quartz crushed at 25° (after a prior heating to 800°C) and the poles to the fracture surfaces upon which they rested.

increased frequency of cleavage directions parallel or sub-parallel to the unit rhombohedral directions as compared with quartz crushed at room temperature without any pre-heating.

The shattering effect of inversion is thus at least partly responsible for the higher rhombohedral peaks of figures 4B and 5 when compared with figure 4A. Prismatic cleavage is not produced by the inversion and the resultant peak in the 88-90° class of figure 6 is close in value to that in figure 4A.

Effect of method of crushing.—A study of quartz fragmented by hammermill action (fig. 7) leads one to believe that the sharp, hammermill blows reduce the frequency of distributive cleavage surfaces and favor sharp, clean

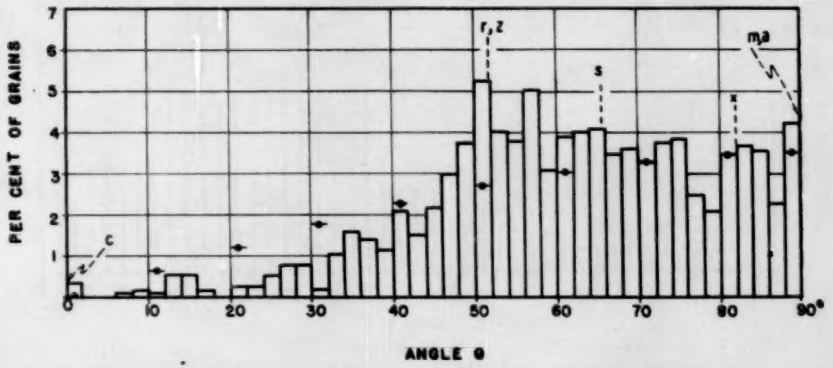

Fig. 7. Histogram showing relation between the *c*-axis of 649 grains of crystal quartz crushed in a hammermill at 25°C and the poles to their fracture surfaces.

breaks more nearly parallel to the preferred cleavage directions. The *r*, *z* peak is now located in the 50-52° class. The secondary peak in the 56-58° class may be fortuitous. Evidence for cleavage parallel to *s* and *x* is perhaps slightly better than in the previous histograms. The prismatic cleavage again shows up well.

Varieties of quartz.—Results for milky and rose quartz are presented in figures 8 and 9. The main characteristics of these distributions are generally similar to those found for rock crystal. It should be noted that the rose quartz, although it contained numerous tiny elongated inclusions aligned parallel its *c*-axis, did not show an abnormally good prismatic cleavage.

SUMMARY AND APPLICATION

Vectorial Nature of Fracture in Quartz

Low Quartz.—For low quartz, cleavage is most readily produced parallel to the *r* or *z* planes and next most readily parallel to the prism *m*.

Cleavage parallel to the basal pinacoid certainly exists but is difficult to compare with the others, primarily because the combined *m* and *a* prisms or the combined *r* and *z* rhombohedra have roughly six times the chance of being favorably oriented with respect to the direction of stress for breakage to occur along one of their component faces.

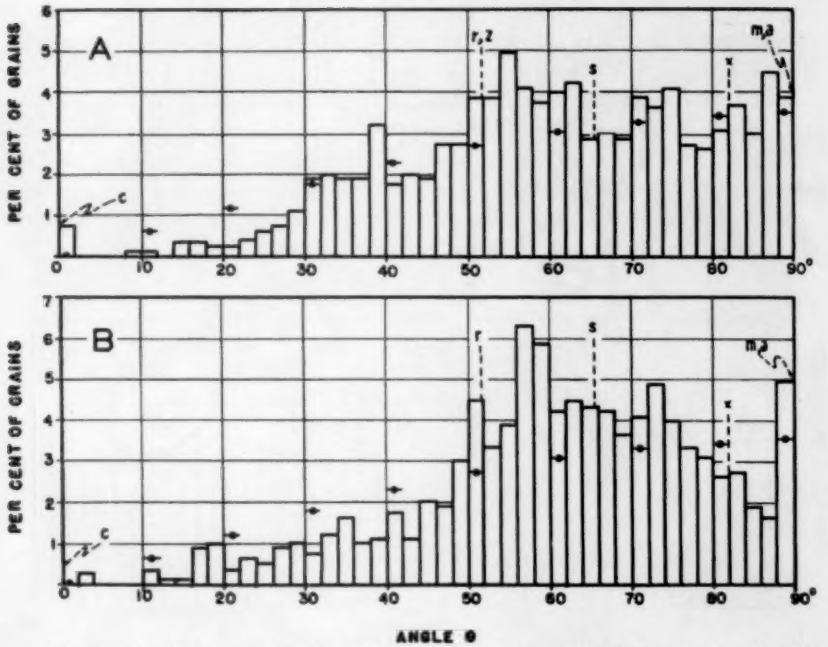

Fig. 8. A. Histogram showing relation between the *c*-axes of 400 grains of milky quartz ramrod-crushed at 25°C and the poles to the fracture surfaces upon which they rested.

B. Same for 400 grains of milky quartz ramrod-crushed at 650-700°C.

Cleavage parallel to the *s* and *x* directions appears possible but present results do not permit a definite conclusion at this time.

Other than the basal pinacoid, quartz shows little tendency to break along planes at high angles to the *c*-axis. This region of strength is generally indicated in the foregoing histograms by the relative sparsity of observations in the 2-36° region.

High Quartz.—The fracture of high quartz is like that of low quartz with the exception that prismatic cleavage is only poorly developed, if at all. The observed relative superiority of cleavage along the 1011 form for high quartz as compared to low quartz may be partly or entirely due to the shatter effect of passage through the inversion temperature.

Relation of Quartz Fracture to Deformation Lamellae

Deformation lamellae in quartz, ably described and discussed by Fairbairn (1941; 1949) and by Ingerson and Tuttle (1945), are most frequently oriented subparallel to the basal pinacoid. Fairbairn (1941, p. 1274) thereby concluded, "Beyond any doubt this property is explainable in terms of its crystal structure . . . ". However, after a re-examination of Fairbairn's Ajibik specimen, Ingerson and Tuttle (1945, p. 99) observed a greater frequency of lamellae at high angles to the pinacoid than had previous investigators. Accordingly, they concluded (p. 102), "Apparently, then, they the lamellae planes

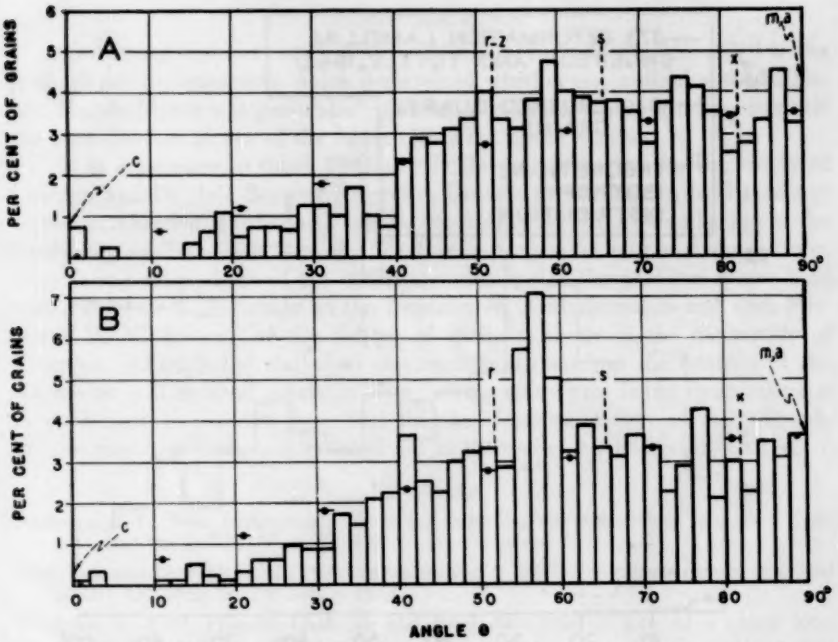

Fig. 9. A. Histogram showing relation between the *c*-axes of 400 grains of rose quartz ramrod-crushed at 25°C and the poles to the fracture surfaces upon which they rested.

B. Same for 400 grains of rose quartz ramrod-crushed at 650-700°C.

are deformation-planes such as would form in homogeneous material. They do, nevertheless, appear to form more readily at low angles to the base."

Comparison of Ingerson and Tuttle's histogram with the writer's results for hammermill-crushed quartz shows a rather antipathetic relationship (fig. 10). The 5-30° range represents planes along which quartz is least likely to fracture yet along which deformation lamellae are most likely to be oriented. Thus, the hypothesis that deformation lamellae represent healed fracture surfaces (already objected to by Fairbairn) seems rather questionable. The fact that lamellae tend to form along planes not prone to fracture also suggests that their positions are, at least in part, crystallographically controlled. This anisotropic control by the crystal structure becomes readily apparent when the theoretical isotropic distribution is compared with that for the deformation lamellae.

CONCLUSIONS

1. For low quartz:
   a. Cleavage is most readily produced parallel to the *r* or *z* planes and next most readily parallel to a prismatic plane whose identity, on the basis of work now in progress, appears to be the prism *m* rather than *a*.
   b. Cleavage parallel to the *s* and *x* directions probably occurs, but present results were not decisive.

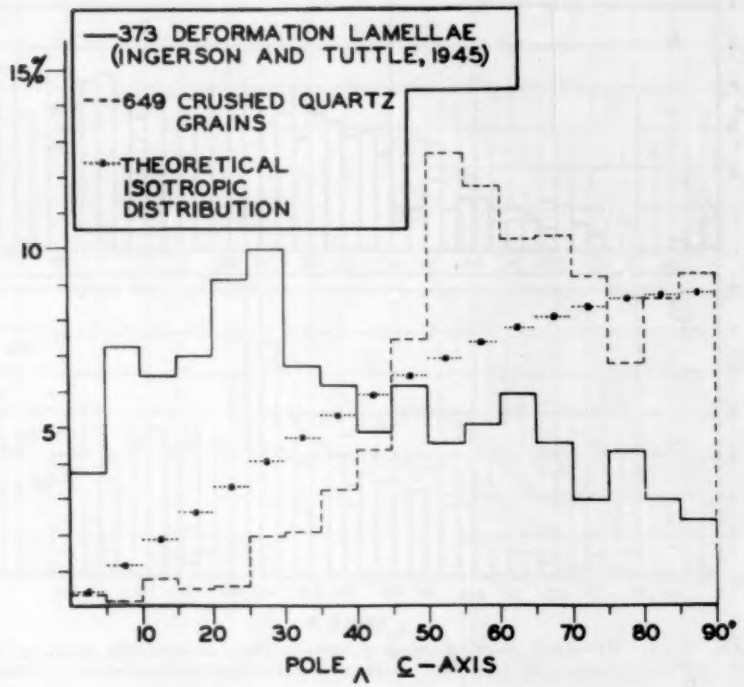

Fig. 10. Histograms (5° class interval) illustrating the antipathetic relationship between the preferred orientations of quartz deformation lamellae and that of quartz fractures as produced at 25°C by hammermill crushing.

c. Cleavage parallel to the basal pinacoid exists but occurs in only ½ to 1 percent of the crushed grains.

d. Fracture along planes making angles of 2° to 36° with the basal pinacoid appears difficult since with few exceptions the isotropic distribution represents greater values than the observed results for this particular region.

2. On going through the inversion temperature 10-16 mesh fragments of quartz evidently developed fractures parallel to the unit rhombohedra, as earlier observed by Wright and Larsen.
3. The cleavage of high quartz is similar to that of low quartz with the following exceptions:

   a. An even greater number of grains of high quartz show cleavage parallel to *r*, perhaps because of the shattering effects of inversion.

   b. Cleavage parallel to the prism is not as well developed as for low quartz.
4. Deformation lamellae in quartz are not likely to represent healed fractures. Their most frequent orientations in quartz coincide with planes along which quartz is least likely to fracture

ACKNOWLEDGMENTS

The writer wishes to express his gratitude to his students who aided in the preliminary investigations which determined whether the method was feasible. Mr. Harold Myers was particularly helpful. Mr. Louis Walter kindly permitted the reproduction of one of the figures from his recent thesis.

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SYNTHESIS OF SPURRITE AND THE REACTION WOLLASTONITE + CALCITE ⇌ SPURRITE + CARBON DIOXIDE*

O. F. TUTTLE and R. I. HARKER

ABSTRACT. A univariant PT curve has been located for the reaction wollastonite + calcite ⇌ spurrite + carbon dioxide. At 5000 pounds per square inch carbon dioxide pressure, the reaction takes place at about 1000°C. Theoretical considerations indicate that tilleyite is expected to have a PT stability field on the low temperature side of the above PT curve. However, tilleyite has not been synthesized and it is suggested that, the accepted composition of tilleyite may be in error, the observed breakdown of spurrite may be metastable, or the observed PT curve may end at a quintuple point involving the five phases wollastonite, calcite, spurrite, tilleyite, and carbon dioxide.

INTRODUCTION

Metamorphism of siliceous carbonate rocks was first considered systematically by Bowen (1940) who proposed that the mineralogical changes which may take place in these rocks during metamorphism were related to a progressive series of reactions of increasing decarbonation. He proposed thirteen steps, involving ten mineral phases, which were related to increasing temperatures of metamorphism and, with certain reservations, each step could be represented by a single PT curve. The PT curves can be used as geological temperature recorders, indicating the grade of metamorphism, if certain conditions are fulfilled. The reaction calcite + wollastonite ⇌ spurrite + carbon dioxide, one of the thirteen steps, will be considered here. ($3\ CaCO_3 + 2\ CaSiO_3 \rightleftarrows 2\ Ca_2SiO_4.CaCO_3 + 2\ CO_2$).

F. E. Wright (1908) first described spurrite in specimens collected by J. E. Spurr at a contact between impure limestone and a diorite, at Velardena, Durango, Mexico where it is associated with gehlenite and hillebrandite. It has subsequently been found elsewhere, and at several localities hand specimens can be collected that consist almost entirely of spurrite.

Spurrite is one of the few silicates which have apparently not previously been synthesized. Wright (1908) reported that an attempt to synthesize spurrite gave orthorhombic crystals of unknown composition. Eitel (1923) reported the synthesis of a similar phase which he believed to be a high temperature modification of spurrite, and placed the inversion at 1200°C at a carbon dioxide pressure of 90 atmospheres. However detailed data were not presented to definitely establish whether the synthetic material is a high temperature polymorph or whether it is a lime silicate.

Spurrite was synthesized reproducibly during our recent study of the system $CaO—MgO—SiO_2—CO_2$, and some information was obtained on its stability.

EXPERIMENTAL METHODS

All experiments have been carried out in cold-seal pressure vessels (Tuttle, 1949; Harker and Tuttle, 1955, p. 210) under controlled and measured pressures and temperatures. Elevated temperatures were obtained by a resistance furnace which surrounds the pressure vessel during the runs, but which can

* Contribution No. 56-5 from the College of Mineral Industries, The Pennsylvania State University.

be removed for rapid quenching. Runs are quenched by immersing the pressure vessels in cold water. A galvanometer type controller, actuated by a thermocouple placed in the furnace above the pressure vessel, controlled temperatures to ± 2°C for short runs and ± 5°C for long runs. Pressures were generated by a pump designed by Roy and Osborn (1952) and modified by Harker and Tuttle (1955). Pressures were measured with a Bourdon gauge which gave readings reproducible to ± 3 percent.

Equilibrium determinations were carried out using as the starting materials either an analytical reagent calcium carbonate and finely ground silica glass, ground together in an agate mortar in the proportion 5 $CaCO_3$:2 SiO_2, or synthetic spurrite obtained by crystallizing some of this mixture. Samples weighing approximately 30 milligrams were used in the experiments. Carbon dioxide was always released at the beginning of the quench to prevent carbonation during quenching. Further details of the general method are given by Harker and Tuttle (1955).

RESULTS

Spurrite was found to break down to wollastonite and calcite with decreasing temperature within the CO_2 pressure range investigated. It can be readily synthesized on the high temperature side of the univariant PT curve shown in figure 1, and it crystallizes readily in the presence of CO_2. (For example, a mixture of calcite and silica glass will crystallize almost completely to spurrite in six days at 950°C and 2,500 pounds per square inch CO_2 pressure.) Unlike the synthetic materials of Shephard (Wright, 1908) and of Eitel (1923) the material produced under the conditions now described appears from its x-ray powder diffraction pattern and optical properties to be identical with natural spurrite. The x-ray powder diffraction patterns of the synthetic spurrite and two natural spurrites (one from the type area) are shown in figure 2. The optical properties compare as follows:

Synthetic	*Natural*
n_α = 1.638 ± .002	1.640 ± .002
n_β* = 1.671 ± .002	1.674 ± .002
n_γ = 1.676 ± .002	1.679 ± .002
γ-α = .038	.039
2v** = 40° ± 5°	40
Sign (–)	(–)
Twinning: polysynthetic	polysynthetic
Dispersion: crossed	crossed

* Calculated from 2v, α and γ.

** Estimated from centered acute bisectrix interference figure.

Pseudowollastonite is commonly produced metastably in experimental studies of systems in which $CaSiO_3$ appears as a phase; however, in this investigation wollastonite was produced in all experiments. X-ray powder diffraction patterns of the synthetic wollastonite, together with pseudowollastonite and natural wollastonite, are shown in figure 3. The intensity differences be-

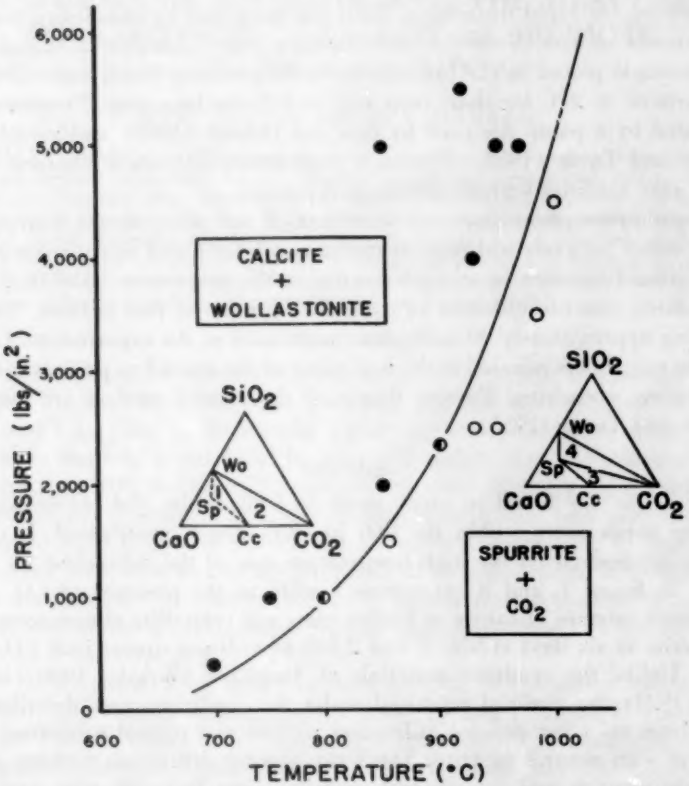

Fig. 1. Pressure (CO_2)-temperature curve for the univariant reaction $3CaCO_3 + 2CaSiO_3 \rightleftarrows 2Ca_2SiO_4.CaCO_3 + CO_2$. Wo = Wollastonite, Cc = Calcite, Sp = Spurrite. The numbers correspond to the assemblages discussed on page 5.

tween the lines of the natural and synthetic wollastonite is a result of orientation in the natural material.

Experimental results are illustrated in figure 1 and are tabulated in table 1. The PT curve represents pressures and temperatures at which the four phases of the reaction calcite + wollastonite $\rightleftarrows$ spurrite + carbon dioxide are stable together. Gibbs phase rule states that P + F = C + 2, where P is the maximum number of phases that can coexist in equilibrium; F is the number of degrees of freedom of the system, and C is the number of components in the system. Thus, when P = 4, F = 1, and for each temperature the four phases calcite, wollastonite, spurrite and CO_2 can coexist in equilibrium at only one pressure. Thus the PT curve of figure 1 indicates the pressure for each temperature at which the four phases can coexist in stable equilibrium. On either side of this curve, P = 3 and F = 2 so that regions of divariancy occur in which the following assemblages can coexist: (1) calcite + wollastonite + spurrite, (2) calcite + wollastonite + CO_2, (3)

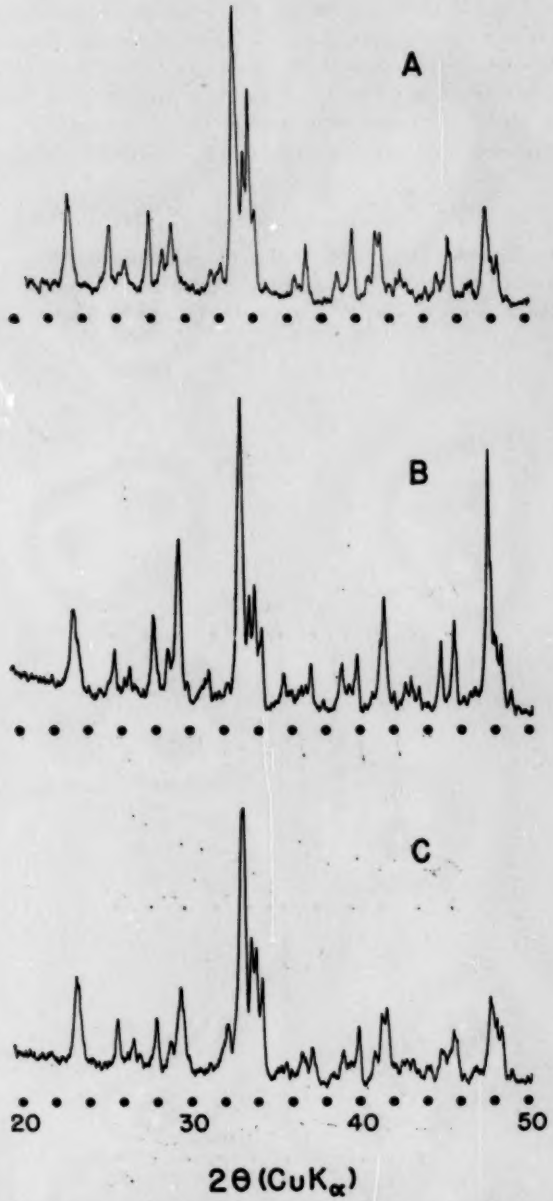

Fig. 2. X-ray powder diffraction patterns of: *A.* Natural spurrite from Velardena, Mexico; *B.* Natural spurrite from Luna Co., New Mexico; *C.* Synthetic spurrite grown at 950°C. and 2,500 lbs./in.^{2}

calcite + spurrite + CO_2, and (4) wollastonite + spurrite + CO_2. Assemblages 1 and 2 can exist stably on the low temperature side of the PT curve while 3 and 4 are stable above the curve. The triangular diagrams on either side of the univariant PT curve in figure 1 depict the range of compositions involved in this reaction. Assemblages are represented on these triangular diagrams by areas outlining compositions involved, and their numbers are shown to emphasize that the assemblages can be readily deduced from such diagrams.

PETROLOGICAL CONSIDERATIONS

The join wollastonite-calcite is the most important feature of the triangular diagrams of figure 1 since these two minerals cannot coexist on the high temperature side of the PT curve. If they are found together with no

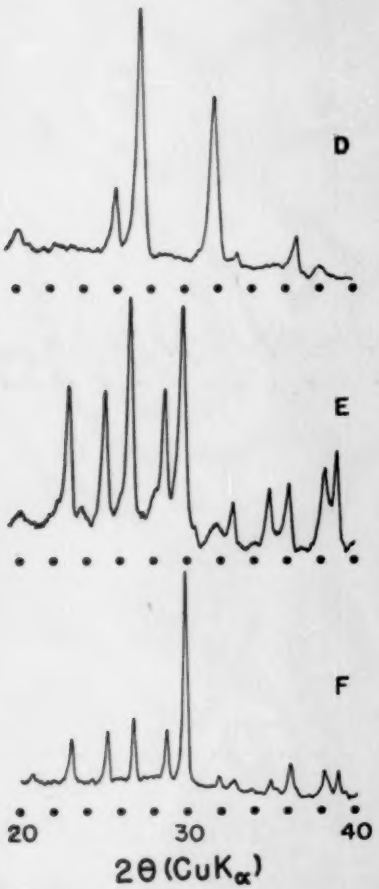

Fig. 3. X-ray powder diffraction patterns of pseudowollastonite(D), natural wollastonite(E), and synthetic wollastonite(F).

associated spurrite, then it can be assumed that the PT conditions on the high temperature side of the curve have not been attained. Calcite, wollastonite and spurrite can coexist on the low temperature side of the PT curve in the absence of CO_2 vapor, but CO_2 is probably always present during the progressive metamorphism of siliceous carbonate rocks as an additional phase, either as a dissociation product of the carbonates or as connate CO_2. Hence unless detrital spurrite is present it is reasonable to expect that the appearance of spurrite indicates that the conditions on the high temperature side of the curve have been attained.

TABLE 1

Experiments on the Reaction

$$3CaCO_3 + 2CaSiO_3 \rightleftarrows 2Ca_2SiO_4.CaCO_3 + 2CO_2$$

Run No.	Initial Material	Temp. °C.	Pressure lbs./in.^{2}	Time Hours	Result
912	CG	920	5500	72	CW
915	S	920	5500	72	CW
895	CG	850	5000	96	CW
896	S	850	5000	96	CW
984	CG	950	5000	144	CW
985	S	950	5000	144	CW
1009	CG	970	5000	72	CW
1010	S	970	5000	72	CW
1084	CG	1000	4500	38	SCW
1085	S	1000	4500	38	S
975	CG	930	4000	80	CW
976	S	930	4000	80	CW
1069	CG	985	3500	48	SCW
1070	S	985	3500	48	S
1017	CG	930	2500	144	SCW
1018	S	930	2500	144	S
874	CG	950	2500	144	S
889	CG	950	2500	48	SCW
890	S	950	2500	12	S
1038	CG	900	2250	240	CW(S)
1039	S	900	2250	240	S(CW)
1125	CG	850	2000	168	CW
1126	S	850	2000	168	CW
1109	CG	850	1500	168	SCW
1110	S	850	1500	168	S
1128	CG	750	1000	432	CW
1129	S	750	1000	432	CW
1143	CG	800	1000	240	CW
1144	S	800	1000	240	S
597	CG	700	400	120	CW

C = calcite
W = wollastonite
S = spurrite
G = silica glass
( ) = trace

It is interesting to note that calcite and quartz are stable together up to 600°C at 5000 pounds per square inch CO_2 pressure (Harker and Tuttle, 1956) while wollastonite and calcite remain stable together up to 1000°C at the same pressure; the 400°C difference in temperature between the appearance of wollastonite and that of spurrite is reflected in the common occurrence of the former at granite-limestone contacts and the latter at diabase-limestone contacts. The rare occurrence of spurrite at granite-limestone contacts is significant because if the pressure of carbon dioxide is low, spurrite should be a common mineral along with wollastonite as temperatures at these contacts are known to be at least as high as 700-750°C. *The fact that spurrite is not common at these contacts indicates that the pressure of carbon dioxide is relatively high and may well be commensurate with the rock load.* This also suggests that the vapor phase is rich in CO_2 and relatively poor in H_2O as the presence of large amounts of H_2O would lower the partial pressure of CO_2 and thereby permit calcite and wollastonite to react to give spurrite.

Another interesting feature of this study is the absence of tilleyite above and below the portion of the spurrite PT curve which has been investigated. In any decarbonation scheme tilleyite is expected to have a stability field on the low temperature side of the spurrite curve. It is expected to develop by the reaction wollastonite + calcite $\rightleftarrows$ tilleyite + carbon dioxide, and the univariant PT curve for this reaction should then be followed at higher temperatures by a curve for the reaction tilleyite $\rightleftarrows$ spurrite + carbon dioxide ($Ca_3Si_2O_7.2CaCO_3 \rightleftarrows 2Ca_2SiO_4.CaCO_3 + CO_2$).

The absence of tilleyite may be the result of failure to obtain equilibrium, the accepted composition of tilleyite may be in error, or the univariant PT curves for the reactions wollastonite + calcite $\rightleftarrows$ tilleyite + carbon dioxide and tilleyite $\rightleftarrows$ spurrite + carbon dioxide may intersect at a quintuple point. Perhaps the first mentioned possibility can be resolved by studying the breakdown of natural tilleyite and we propose to attempt this as soon as suitable material is obtained. Two analyses of tilleyite reported by Nockolds (1947) and one by Larsen and Dunham (1933) show more than 1 percent water, and although this may represent hydrated alteration products, it is suggested that water may be essential in the structure of tilleyite. The fact that tilleyite is commonly secondary after spurrite also suggests a hydrated mineral, although it is not impossible that tilleyite is replacing spurrite and wollastonite as a result of retrograde metamorphism in the presence of excess CO_2. There are two possible sets of reactions if the determined PT curve terminates at a quintuple point, one in which tilleyite is unstable at high pressures in the presence of carbon dioxide, and a second in which tilleyite is unstable at low pressures of carbon dioxide. The following assemblages appear in each of these reactions: (1) wollastonite, calcite, tilleyite, carbon dioxide; (2) wollastonite, calcite, spurrite, carbon dioxide; (3) wollastonite, calcite, spurrite, tilleyite; (4) wollastonite, spurrite, tilleyite, carbon dioxide; and (5) calcite, spurrite, tilleyite, carbon dioxide. As spurrite, tilleyite and carbon dioxide have compositions which lie on a line in the ternary system $CaO—SiO_2—CO_2$ the last two reactions (4 and 5) involve binary equilibria and these two as-

semblages can be reduced to the reaction tilleyite ⇄ spurrite + carbon dioxide.

The first mentioned possible quintuple point is illustrated schematically in figure 4A. Reactions which take place along each of the four curves are shown together with the compositions involved. It can be seen that tilleyite

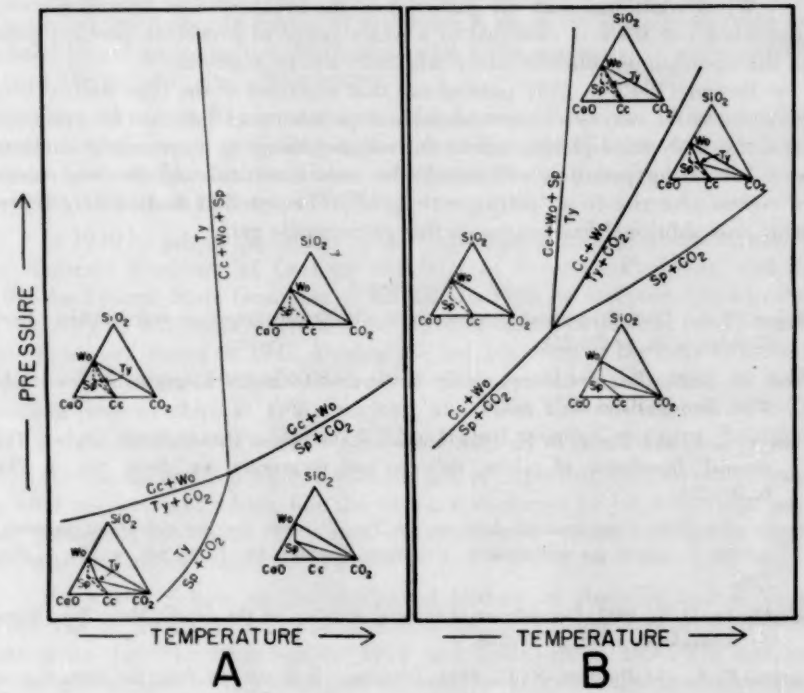

Fig. 4. Hypothetical P-T diagrams illustrating the two possible arrangements of P-T curves about a quintuple point. In the presence of excess CO_2 tilleyite can only be stable at pressures below the quintuple point in 4A, and above the quintuple point in 4B. Wo = Wollastonite; Ty = Tilleyite; Cc = Calcite, and Sp = Spurrite.

can exist only at low carbon dioxide pressures with this arrangement of curves. This quintuple point was proposed by Korzhinsky (1937) as one of a series of changes that may take place at depth as a result of increasing pressure. The second possible quintuple point is illustrated in figure 4B and here tilleyite would have a minimum pressure at which it is stable. In each instance the nearly vertical curve involves only solid phases and to realize this reaction it would be necessary to work in the "vapor deficient" region. Either of the quintuple points described above would explain the fact that spurrite is of more common occurrence than tilleyite because spurrite could develop from calcite + wollastonite in the presence of excess CO_2 throughout the entire range of pressures shown, whereas the pressures at which tilleyite could de-

velop from calcite + wollastonite in the presence of excess CO_2 would be limited by the quintuple point.

If the first of the above quintuple points (fig. 4A) represents the true relations, the pressure at the invariant point must be very low indeed, perhaps less than 500 pounds per square inch. Only those contacts formed at very shallow depths would be expected to carry tilleyite under such circumstances.

When additional data are gathered on the breakdown of natural tilleyite, and when our work is extended to a wider range of pressures, perhaps some of the uncertainties introduced by this study will be resolved.

Bowen (1940, p. 273) pointed out that reactions of the type studied here give rise to PT curves with considerable slope whereas PT curves for reactions involving only solid phases, where the volume change is more nearly constant with increasing pressure, will usually be nearly vertical and the two classes of curves give rise to a "petrogenetic grid". The reaction studied here represents one additional fixed curve on this petrogenetic grid.

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In Memoriam

WILLIAM HENRY TWENHOFEL
1875-1957

We record, with a deep sense of loss, the death of William Henry Twenhofel, an Associate Editor of this Journal since 1935. He succumbed to a brief illness on January 4, 1957, after a long and uncommonly active career as teacher, scientist, and administrator.

Born in humble circumstances on a farm near Covington, Kentucky, he taught in the country and village schools until he could afford to come to Yale where he earned his way and received the degrees of B.A., M.A., and Ph.D. in 1908, 1910, and 1912 respectively.

In 1910 he joined the faculty of the University of Kansas where he served as Assistant Professor of Geology and later as Associate Professor, and in 1915 he became State Geologist of Kansas. In 1916 he accepted a call to the University of Wisconsin where he held a professorship in geology for 29 years before his retirement in 1945. During the last five years of his term he served also as Chairman of the Department of Geology. Since retirement he held visiting professorships at Yale, Stanford, and Tulsa Universities, and in 1947 he received the honorary degree of D.Sc. from Louvain University, Belgium.

He was an inspired teacher with the gift of imparting his own enthusiasm to students, many of whom, like the writer were drawn by his magnetism into the geological profession. To a host of friends and former students he came to be known affectionately as "Twen".

He was a Fellow of the Geological Society of America and a Vice-President in 1930. He was also a Fellow of the Paleontological Society of which he was Vice-President in 1924 and President in 1930. He was an honorary member of the American Association of Petroleum Geologists and of the Society of Economic Paleontologists and Mineralogists of which he was President in 1935. In 1952 he received an honorary award from the Society of Exploration Geophysicists. He was chairman of the Division of Geology of the National Research Council from 1931 to 1934.

His many publications include *The Geology of Anticosti Island*, *Invertebrate Paleontology* (with R. R. Shrock), *Principles of Sedimentation*, and the monumental *Treatise on Sedimentation* of which he was editor and collaborator.

In 1924 Twenhofel organized the Committee on Sedimentation of the National Research Council of which he served as chairman from 1924 to 1931. He was also a moving spirit in launching the Journal of Sedimentary Petrology, and served as its editor from 1933 to 1950. In thus pioneering in, and stimulating the study of sedimentary processes he has earned an enviable position among the leaders of American geology.

The Direction of Time; by HANS REICHENBACH. P. xi, 280; 39 figs. Berkeley, 1956 (University of California Press, $5.50).—Physical science is pervaded by a strange paradox: every normal person carries within himself an intuitive and unambiguous sense of the direction of time; yet all basic physical laws are reversible and therefore provide no facility for defining a direction of time. Any elementary phenomenon can be converted into another possible elementary phenomenon merely by changing the sign of the time variable; a ball bouncing upward is nothing more than a falling ball with the time variable reversed. Why then is it that we, when observing a ball which periodically falls and rebounds, believe to perceive a sequence of different motions within uniformly flowing time, instead of seeing the same continuing motion on a time scale which periodically changes its direction?

Needless to say, this problem has received a great deal of attention from scientists as well as philosophers, the latter having done about as much to becloud as to solve it. The scientific aspects of it are now entirely clear in the sense in which scientists demand clarity, and the last important contributions to its solution were made by Boltzmann, Poincaré, Ehrenfest and Von Neumann. But a problem may be solved scientifically, in accordance with the standards of rigor of the time, and still have important philosophic implications.

Reichenbach's book is not primarily devoted to these philosophic implications as one might at first be led to expect. It offers an explanation, and a very sober one, of the scientific situation surrounding the problem of reversibility. As I see it, the book attempts two things: 1) a careful presentation of the arguments of Boltzmann and Ehrenfest in simple terms and with an unusual measure of logical scrutiny; 2) a restatement of these arguments in a language which Reichenbach himself developed, a language meaningful within the framework of his own theory of probability. In my opinion the presentation has suffered because of the author's insistence on his private formulation, which lacks the elegance of more usual mathematical treatments.

The resolution of the paradox, stated in a manner much cruder than Reichenbach's, is this. Every *isolated* system has a reversible history, and its entropy rises and falls. By consideration of a single isolated system one would arrive neither at the second law of thermodynamics nor at a unique direction of time. But an *ensemble* of isolated systems, suitably withdrawn from the universe, exhibits a totality of behavior within which the second law has meaning and a time direction can be defined. The theory of ensembles of time series, now fully developed, allows a very succinct analysis of this state of affairs which would have taken far less space than Reichenbach devotes to it. But it would not have exposed all the detail which the present treatment uncovers.

The book is posthumous and has not had the finishing touches of the author's able hand. It belongs to the best of his writings; I would rate it just below his *Philosophie der Raum-Zeitlehre*, higher than his other books. It achieves the two purposes it sets out to accomplish in admirable fashion, and will enrich the reader's memory with examples and phrases that are not easily forgotten. Causal net, genidentity, branch system, hypothesis of branch

structure, principle of the parallelism of entropy increase, the principle of the common cause, are typical coinages of Reichenbach's inventive mind. They may not always clarify, but they incite thought partly because of their awkwardness. There are places where usage is erroneous ("saddle point") and others where difficulties have been overlooked, (In fig. 20, how can the entropy of the universe equal that of an isolated system, a situation depicted at the branch points?) Surely such imperfections, even if minor, would have been removed by the author had he been permitted to revise this work. Despite its faults the book stands as a monument to the careful, never-tiring, ever-wary empiricism of Hans Reichenbach, whose passing innumerable friends, be they critics or admirers, mourn with genuine regret.

HENRY MARGENAU

The History of the British Flora. A factual basis for biogeography; by H. Godwin. P. viii, 384; 119 figs., 26 pls., 1 tbl. New York, 1956 (Cambridge University Press, $16.60).—This book falls naturally into two parts. One is an annotated compilation of the stratigraphic occurrence of plant remains in Great Britain from the end of Pliocene time to the present. This, the factual basis of biogeography of the subtitle, occupies over half the volume. Fossils are listed by species and a summary account is given of the collector, locality, type of fossil and stratigraphic range. In addition, most species are accompanied by a critical commentary one paragraph to several pages in length.

The other part of the book consists of background information about methods and Pleistocene environmental changes and a concluding essay on entry, expansion, contraction and extinction of the plants that have contributed to the British flora.

The treatment of documented plant fossils is excellent. A very large proportion of the records, and almost all the well-dated ones, were provided by Godwin and his students, and biogeographers of the British Isles are very greatly in his debt not only for having brought this scattered information together in so compact a form but also for his careful and highly sophisticated evaluation of the fossil evidence.

Although no effort is made to supplant existing palynological textbooks much of the information about laboratory methods, fossil identification and the zoning and interpretation of pollen diagrams will be very useful to students of Pleistocene vegetation in all parts of the world. In particular, many could profit by emulation of the way in which Godwin examines and rejects the evidence for a xerothermic interval in Britain. The essay on floral changes is something of a landmark in the development of biogeography, for it brings to bear on many of the classic problems an unparalleled wealth of stratigraphic evidence. Nevertheless this part of the book lacks dramatic impact, for the major conclusions have been put forward in a number of short publications during the past seven years. Many of them are already part of the working knowledge of alert biogeographers and it may be doubted whether biogeographers of other kinds, having ignored the preliminary notes, will take the trouble to digest this book.

The treatment of broad aspects of Pleistocene history is less satisfactory. Those who anticipated a detailed statement of Godwin's views on British glacial stratigraphy will be disappointed, for he has been extremely conservative in his interpretation of the geological evidence. The account of radiocarbon dating might have been written in 1952, and the few carbon dates included in the book are early ones of Libby, obtained by the screen-wall technique. One cannot suppress a twinge of regret that publication was not delayed a few years to include the results of radiocarbon dating of the last glaciation in the British Isles. Perhaps we may hope for another volume that will cover the Pleistocene history in a more adequate way.

In style and format the book is pleasantly legible. Typographical errors are very few. The figures are clear and attractive, as are the half-tone plates of macroscopic subjects, but some of the photographs of seeds and pollen grains lack definition.

This is a book no paleoecologist, biogeographer or glacial geologist can afford to ignore. Unfortunately it is also one few can afford to buy.

D. A. LIVINGSTONE

Some corrections in mechanical analysis by pipette method; by S. I. MALININ, Moskov. Obshch. Ispytateley Prirody Byull., Otdel Geol., tom XXIX, vypusk 2, p. 101, 1954.—As pointed out by Krumbein and Pettijohn (1938, p. 163, *Manual of Sedimentary Petrography*, Appleton-Century-Crofts, Inc., New York), the pipette method was developed independently in 1922-1923 by Robinson in England, Jennings, Thomas, and Gardner in the United States, and Krauss in Germany. Malinin (1954), however, claims that the method was first developed by V. G. Glushkov in 1911, and was adopted for the mechanical analysis of recent marine sediments by the Government Oceanographic Institute in 1931.

M. V. Klenova (1933, in Malinin, 1954) first noticed that the results obtained by the pipette method differ from those obtained by Osborn's method, which specifies control of the grain size of various fractions by the microscope. Other Soviet investigators confirmed Klenova's findings, and it was determined that on using the pipette method the amount of coarse fraction is lower and the amount of fine fraction is higher than by the accurate methods involving the control of grain size under the microscope. The total amount of pelitic material is also determined inaccurately on using the pipette method.

A comprehensive study has been undertaken by Malinin in order to determine the accuracy of pipette method. Drops of the suspension were collected from a fixed depth at the end of different intervals of time (starting at 20 seconds after mixing) and examined under the microscope. The results of his investigation led him to believe that Stokes' formula is completely unfit for polymineral dispersions. The following time intervals for the collection of samples are proposed by Malinin in order to obtain correct results.

1. 0.01-0.05 mm fraction—69 seconds after shaking the suspension, from a depth of 150 mm from the initial level.

2. 0.005-0.01 mm fraction—29 minutes after shaking, from a depth of 150 mm.

3. 0.001-0.005 mm fraction—8 hours after shaking for the sediments containing more than 50% of pelitic material, and 7 hours 30 minutes for sediments having less than 50% of pelitic material.

4. <0.001 mm—24 hours after shaking the suspension. The latter two fractions are taken from a depth of 100 mm from the initial level of suspension.

Inasmuch as the findings of the reviewer (Tchillingarian, 1952, *Jour. Sedimentary Petrology*, v. 22, no. 4, p. 229-233) also showed that different results are obtained on using different dispersing agents, the accuracy of the pipette method should be thoroughly investigated.

GEORGE V. CHILINGAR

Contribution à l'Etude des Minéraux Secondaires d'Uranium Français; by JEAN CHERVET and GEORGES BRANCHE. P. 190, 35 figs., 9 pls. Sciences de la Terre, vol. 3, nos. 1-2. Ecole Nationale Supérieure de Géologie. Nancy, 1956 for 1955.—The second of this series of monographs on the uranium deposits of metropolitan France and the French dominions is concerned with supergene mineralization. The present handsomely produced text is primarily mineralogical, comprising detailed descriptions of some thirty secondary uranium minerals found in the mines of France, together with a critical review of the literature on species yet unrecognized from French territory. Among the illustrations there are twenty-four in color, from photographs by M. Destas, which are perhaps the finest reproductions of mineral specimens that have ever been printed. A second paper in the same volume (p. 193-235), by M. André Poughon and M. Marcel Moreau, gives a full account of the deposits of parsonsite at Lachaux, near Vichy, the only field of this rather rare uranium mineral that has yet been exploited commercially.

C. F. DAVIDSON

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